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MILTON WHITNEY, Chief.

STUDIES ON THE MOVEMENT OF SOIL MOISTURE.

BY

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LETTER OF TRANSMITTAL.

U. S. Department of Agriculture,
Bureau of Soils,
Washington, D. C., November 27, 1906.

Sir: I have the honor to transmit, and to recommend for publication as Bulletin No. 38 of this Bureau, the manuscript of a technical paper entitled "Studies on the Movement of Soil Moisture."

Respectfully,

Milton Whitney,
Chief of Bureau.

Hon. James Wilson,
Secretary of Agriculture.

PREFACE.

The movement of the water in the soil, whether by the capillary forces exerted in the finer soil spaces at soil surfaces, or by distillation from one point to another, has long been regarded as a very important phenomenon and has often been investigated. Nearly all these studies, however, have been concerned with the movement of free water, in practically unlimited amounts, into a more or less dry soil. The much more practical problem of the movement of water from a soil short of optimum content to an even drier portion, or the distribution of the moisture content of the soil under conditions simulating those existing in the field, has hitherto received much less attention. It has become evident, from the studies which have been made so far, that the movement of moisture in the soil under such conditions is very slow, much slower as a rule than the growth of roots, which can, in a sense, reach out for needed water supplies more rapidly than the soil can bring the water to them. But while this feature of the problem can now be safely held to have but little if any immediate practical importance, the problem is one of great significance in the long run, especially in handling soils subject to conditions of drought.

In the present bulletin it is shown that the loss of water by evaporation from points below the surface, while it does take place in measurable quantities, is nevertheless quite small, and is negligible in comparison with the losses taking place at or very near the surface. The movement of water vapor through the soil is shown to follow the law governing the diffusion of other gases through porous media, and is quite slow. Mulching decreases or inhibits the capillary flow, and diffusion through the mulch is practically negligible. This practice is very effective in conserving soil moisture, and is founded on sound scientific principles. An especially interesting illustration is brought out in the comparison of the loss of water from a soil under arid and humid conditions, respectively. As might be expected, the loss at first is much more rapid under the arid conditions, so rapid in fact as to overtax the soil's ability to move water from within to the surface by capillarity, and in consequence a dry layer or mulch is formed which keeps the subsequent losses far below those which take place from the soil under humid conditions, where the capillary flow

to the surface persists until the moisture content of the whole soil is very low indeed. These laboratory experiments, therefore, clear up in a very satisfactory manner the well-known and apparently contradictory facts observed in the field that the soils of arid regions, at depths a little below the surface, are generally wetter and hold their moisture for much longer periods than do the soils of humid areas in dry seasons.

An examination of the curves representing the distribution of moisture in the soil has suggested that if the subject be regarded from the standpoint of dynamical equilibria phenomena there appear certain analogies to the theory of electrical and thermal potential. This is a fact of great importance in the theoretical study of the subject, since we are in possession of a well-developed theory of electrical and thermal potential which can be applied to a considerable extent to the phenomena of soil moisture; it is of probable practical value, because it suggests and gives direction to further experiments and serves as a basis for the correlation of a large number of observations already made, but having hitherto a local rather than a general value.

It is clearly recognized that the analogy is imperfect in that the capillary potential and resistance to flow are dependent upon the moisture content of the soil, whereas electrical and thermal resistance are practically independent of the amount of current and heat passing. Nevertheless this point of view and method of attacking the problem promise to become very helpful, and are believed to be a distinct advance in the treatment of a subject which has hitherto suffered on account of serious experimental difficulties and the lack of a coherent guiding theory.

FRANK K. CAMERON.

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STUDIES ON THE MOVEMENT OF SOIL MOISTURE.

A.—THE LOSS OF SOIL MOISTURE BY DIRECT EVAPORATION FROM POINTS BELOW THE SURFACE.

When a moist soil dries by exposure to the air over its surface the greater part of the evaporation into the air takes place at or very near the surface of the soil, the water moving up from below toward the surface by capillary action. This is shown by the fact that a dry mulch in which there is little or no capillary flow of water protects the soil, to a great extent, from drying out. It is obvious, however, that unless the pores of the soil are completely filled with water, or unless the soil is completely air dry, there will be some evaporation of the water held in the soil into the air in the pores, and that there will be a gradual escape of this vapor by diffusion to the outside air, except in the very unusual case of the outside air being completely saturated with moisture, when no drying of any sort will occur. In general, therefore, there is a certain quantity of water lost from soils by this direct evaporation from points below the surface, quite independently of capillary motion. The object of the experiments[a] to be described was to get an idea of how much this loss amounted to in certain particular cases.

The general method employed was to expose a surface of water or moist soil to evaporation into a confined space which was in communication with the outside air through a column of soil of known height and cross section. Whatever water escaped from below the column of dry soil had thus to escape by pure diffusion, because there was no capillary connection between the column of soil and the water or moist soil from which the loss was being measured.

In the first experiment the soil from which the evaporation was to take place was spread in a loose layer about half an inch thick in the bottom of a glass tumbler, while the upper 2 inches of the tumbler were filled with coarse dry sand supported on wire gauze and cheese cloth. There was thus a free space of about half an inch left between the damp soil in the bottom and the dry sand above it, so that capillary communication was cut off and whatever loss of water occurred had

[a] The experiments described in this bulletin were carried out by Messrs. J. O. Belz and J. R. McLane.

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to take place solely by evaporation into the free space below the layer of sand and diffusion of the resulting vapor up through the sand to the outside air.

The soil used was a sandy loam from Takoma Park, Md. In three tumblers this soil was mixed with solid manure and then wet up to about 17, 20, and 27 per cent of moisture, taking the weight of the dry material as the basis of computation. In three more tumblers the soil was wet to about 10, 16, and 22 per cent with liquid manure; and in three more the soil was treated with tap water only, the percentages of moisture being in this case 12, 17, and 23. In the bottom of a tenth tumbler water alone, without soil, was placed.

These ten tumblers were placed in a circular row on a turning table in an otherwise unused room at ordinary laboratory temperature. The experiment was performed in December, 1904. For thirty-nine of the ninety-eight hours' duration of the experiment, a breeze of about 3 miles per hour from an electric fan was kept blowing over the tumblers on the turning table, so that they were all under similar conditions as regards evaporation. During the remainder of the time—at night—when the fan and the turntable were at rest, the evaporation took place into still air, but also under similar conditions for all the tumblers. The tumblers were weighed at the start, at the end, and at several intermediate times, and the losses of weight found by difference. The results may be summarized as follows:

(*a*) There were no systematic differences in the losses, connected with the differing water content of the soils.

(*b*) The soils treated with solid manure lost water slightly faster than the others, but only to an extent which might be accounted for by assuming that the heating effect of the rotting manure kept their temperature about 1° C. higher than the others, thus increasing the vapor pressure of the water over them.

(*c*) With this exception the loss of water was the same, within the experimental errors, for all the tumblers, including the one in which the evaporation was from a free water surface, instead of from water distributed in thin films over the surfaces of the soil grains.

(*d*) Comparison of the rates of loss, with and without the fan and turntable in motion, showed that the rate of loss was about three times as great into a 3-mile breeze as into the still air of the room.

(*e*) The rate of loss observed was approximately the same as might have been computed a priori from known data and known principles with the aid of very probable assumptions. The known data referred to are the thickness, cross section, and porosity of the layers of dry sand; the temperature of the experiment, and the humidity of the outside air (both known only approximately); and finally the free diffusion constant of air and water vapor. The known principles are the laws of the free diffusion of gases, and the fact, established by the writer's

experiments,[a] that the rate of interdiffusion of air and carbonic acid through four soils of varying types and conditions was approximately proportional to the square of the porosity[b] of the soil. The assumptions involved are that this rule holds also for the interdiffusion of water vapor and air, that the free space below the sand was sensibly saturated with water vapor, and that the 3-mile breeze was sufficient to prevent any sensible increase in humidity in the air immediately over the tumblers.

From these experimental results the following conclusions may be drawn:

(I) The free-air space below the sand was in all cases equally moist, probably very nearly saturated. In other words, the varying treatment of the soil as regards both manuring and moisture content did not influence the partial pressure of the water vapor in the air space above it.

(II) Hence, when the resistance to escape by diffusion was as great as that offered by a 2-inch layer of very coarse dry sand, no differences were perceptible, if any such existed at all, between the ease of evaporation from the damp soils treated in various ways and from a free water surface.

(III) As was to be expected, the diffusion of water vapor into the air through soils followed the same laws as the diffusion of air and carbonic acid under similar circumstances.

(IV) The actual mean rate of loss of water was equivalent to about 1.4 inches of rainfall per year into still air, and to about 4.3 inches into a breeze of 3 miles per hour, the temperature of the soils and the air having been about 20° to 22° C., and the dew-point about at the freezing point.

In a second experiment a test was made of the rate of escape of water by direct evaporation, without capillary flow, from below layers of moist soil 12 inches thick. Six cylinders, 12 inches high and 3 inches in diameter, were filled with the sandy loam soil already mentioned, wet to about 10 per cent with water. In two cylinders the soil was tamped down hard, so that its porosity at the start, before any of the water had evaporated, was $S=0.436$. In two more the soil was tamped lightly, giving it an initial porosity $S=0.498$. In the remaining two cylinders the soil was loose, having an initial porosity $S=0.520$. The bottom of each cylinder was formed of wire gauze and cheese cloth.

Detachable brass cups were fitted over the bottoms of the cylinders with nearly air-tight joints through which no sensible diffusion of water vapor could occur. The cups contained water, but not enough

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[a] Bul. No. 25, Bureau of Soils, U. S. Dept. of Agr.

[b] By the "porosity" is meant that fraction of the total volume of the soil which is not occupied by either water or solid soil grains.

to come in contact with the soil supported above on the wire gauze and cheese cloth. There was thus a free air space between the water and the soil, so that whatever loss took place from the cups was by evaporation from the water, followed by diffusion of the vapor into the soil, no capillary action occurring except in the soil itself.

These six cylinders were set in a circle on the turntable, and air from the fan kept blowing over them, the motion of the fan and the turntable being, with insignificant exceptions, continuous during the whole fifty-three days of the experiment, which was started on January 20, 1905. At intervals of from four to ten days the soil cylinders and the cups were weighed separately, and at the end the soil was taken out in layers and moisture determinations were made, so that the final distribution of water might be known. We are interested at present only in the losses of water from the cups.

This experiment is not quite so simple as the one described before, because the layer of soil is not, as the sand was, of very nearly constant porosity and wetness, but is gradually drying out, so that the amount of free pore space in it is increasing. Suppose, however, that we consider only what happened during the last seventeen days, when the soil had reached a state of only slow change as regards the distribution of water in it, and compare the rates of loss from the cups with the final mean porosity of the lower 3 inches of soil. It was found that the loss from the cups was nearly proportional to the square of the porosity, the mean variation from exact proportionality being only 1.8 per cent of the mean value of the quantities compared. The results showed a slight run, indicating that the rule was not exactly followed. By assuming that the diffusion was proportional, not to the square, but to the 2.2 power of the porosity, the run vanished and the mean variation from proportionality fell to 1.4 per cent. It thus appears that the rule that the rate of diffusion of water vapor into air through a given soil is proportional to the square of the porosity of the soil is again approximately confirmed.

Computations similar to those already mentioned in connection with the tumbler experiments, and based on the assumption that the partial pressure of the water vapor in the pores of the soil varied uniformly all the way from top to bottom, gave a value about four times as great as the loss from the cups as actually observed. This indicates simply that the partial pressure gradient was not uniform, but much smaller at the bottom than on the average. This is natural enough. The air in a damp soil must everywhere at a depth of several inches be nearly saturated, so that the variations of humidity from point to point must be much smaller than close to the surface.

It is interesting to find that the "porosity square" rule is valid, but the important result of the experiment is the actual amount of the loss from the cups. This was, of course, greatest for the most porous or

loosely packed soil, and also somewhat more rapid as the porosity of the soils increased by their drying out. But the highest rate of loss obtained for the last seventeen days was equivalent to considerably less than one-fifth of an inch of rain per year, an amount which, from an agricultural standpoint, is altogether insignificant.

During this experiment the soil itself was, of course, drying out, and it is worth while to note a few facts regarding this loss of water by the soil. In the first place, it was found that the total loss of water was very nearly proportional to the average water content of the soil, expressed not in per cent of the dry weight, but in grams per cubic centimeter. The percentage of water was in fact very nearly the same for all the cylinders, whereas on account of the different degrees of compactness the water per cubic centimeter differed considerably. The results are shown in Table I.

TABLE I.—*Loss of water from soil of different porosities contained in cups.*

	I.	II.	III.
Mean porosity of the soil	0.47	0.53	0.55
Mean percentage of water during the whole experiment	7.2	7.2	7.3
Mean rate of drying in inches of rain per year	5.24	4.68	4.39
Mean weight of water in the soil in grams	117.9	105.6	100.2
Ratio of rate of drying to mean water content	0.0444	0.0443	0.0439

The distribution of water in the soil columns at the end of the experiment is shown by the curves of figure 1. We see from these curves that:

(*a*) All the cylinders had about 0.4 to 0.5 per cent of moisture in the top inch of soil.

(*b*) Upon going down into the soil the water content increased more rapidly in the compact than in the looser soils.

(*c*) The compact soil had a much more uniform final distribution of water, having lost more from the lower layers and less from the upper, showing, as was to have been expected, a decidedly stronger capillary flow in the compact than in the loose soil.

In a third experiment the soil used was air-dry Cecil clay, with an initial water content of about 1.9 per cent (from 1.8 to 2.1). The same 12-inch cylinders were used as in the last experiment and the general method was the same. As before, the six cylinders were arranged in three pairs of duplicates and the results given are mean values for two cylinders. The initial porosities of the soil were 0.46, 0.57, and 0.66. The medium and loose soils acted very much alike, and after two hundred and sixty-eight days the soil of medium compactness was omitted from the experiment. The experiment ran for four hundred and forty-one days, or from April 17, 1905, to July 2, 1906. The losses of water from the cups below the soil, and the total losses of the whole system, were determined from time to time by weighing, more

water being added to the cups when necessary. The results may be seen from figure 2.

The abscissas represent the number of days since the experiment was started, the actual dates being also marked. The ordinates represent grams of water lost. To convert grams of water into inches of rain the ordinates should be multiplied by 0.00864.

The curves A and B show the total amounts of water lost from the cups. This water has evaporated and passed up in the form of vapor into or through the 12-inch column of soil above, between the water and the outside air of the room. Curves C and D show the simultaneous losses of weight of the whole system, cups and soil together. These curves lie everywhere below the curves A and B, which means that of the total amount of water lost by the cups only a part escaped to

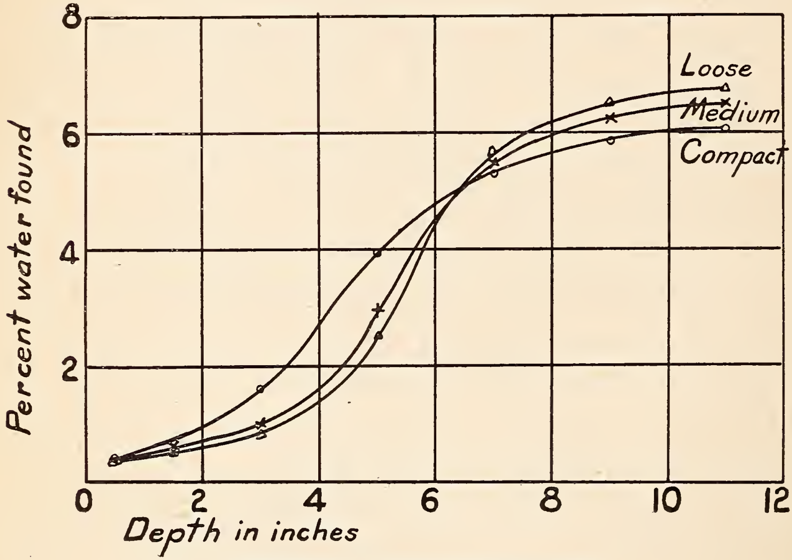

FIG. 1.—Distribution of water in soil after evaporation.

the free air, while a part was taken up by the soil, which thus gained in water content. The amount of water gained by the soil is evidently the difference between the loss from the cups and the loss from the whole system. These differences are shown by curves E and F, which therefore represent the way in which the number of grams of water in the 12-inch columns of soil, 3 inches in diameter, increased as time went on, by reason of the contact of the soil with the saturated water vapor below it. The experimental points plotted are in each case the means for two duplicate cylinders, cylinders Nos. 1 and 2 having an initial mean porosity of 0.46, and cylinders Nos. 5 and 6 having an initial mean porosity of 0.66. A good many more points were obtained than are plotted. The ones plotted were selected so as to be as uniformly distributed as possible and without regard to the actual values.

On examining the figure we see that during the first five or six months, or from the start in April to the beginning of October, the soils continued taking up water. From that time on till the middle of January, 1906, while the average absolute humidity of the outside air was decreasing, the soils lost water. During the warm, damp weather of February they gained again. During the colder weather

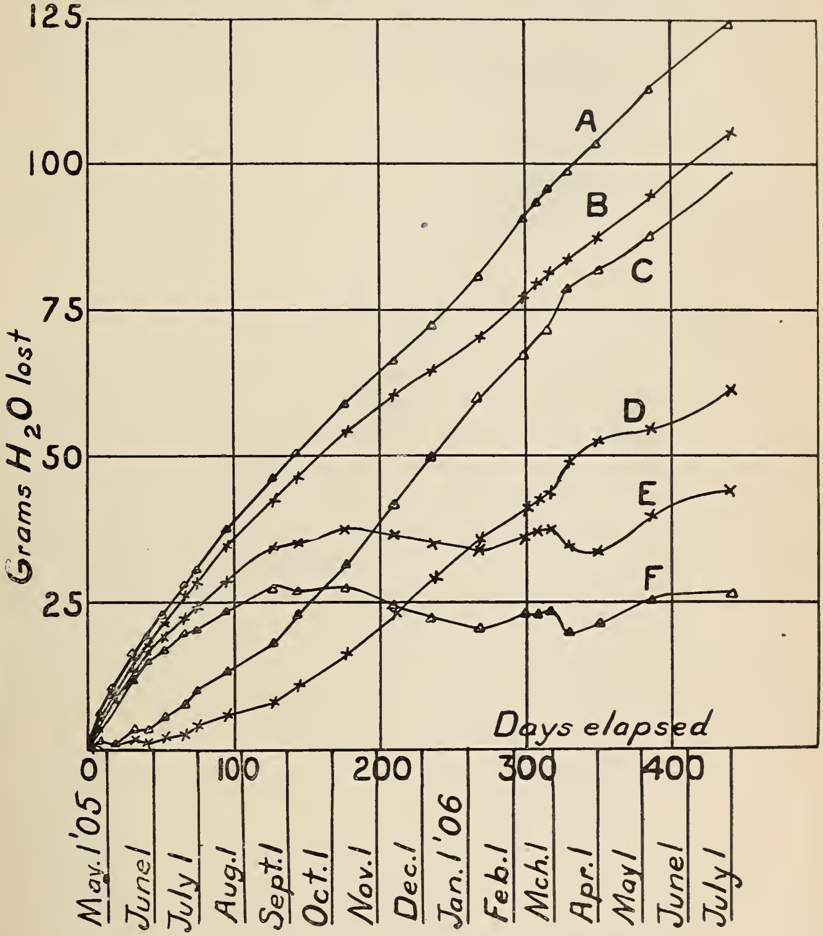

FIG. 2.—Loss of water from below 12-inch layers of soil.

of March they lost once more, while the warmer summer weather of June made them take up water once more. If the experiment were kept running indefinitely similar changes would probably repeat themselves, the average water content staying about the same from year to year. The maximum water contents of the soils, reached about October 1, 1905, were 3.7 per cent for the compact soil (cylinders Nos. 1 and 2) and 4.3 per cent for the loose soil (cylinders Nos. 5 and 6), the corresponding initial water contents having been 1.8 per cent and 2.1

per cent. This phase of the experiment illustrates the slowness with which such a soil comes to equilibrium with the water vapor of the air and the indefiniteness of the term "air-dry."

The fact that after six months curves E and F remain approximately horizontal corresponds to the fact that curves C and D become approximately parallel to the accompanying curves A and B; in other words, that when the soil has ceased to take up much water the additional water lost from the cups (curves A and B) is the same in amount as that lost from the whole system.

Furthermore, we see that from about October 1, 1905, onward the curves A and B are nearly straight lines, or the rates of loss of water from the cups were nearly constant. The average rates of loss, found by taking the slopes of straight lines coinciding as nearly as possible with the later parts of these curves, were:

0.60 inch of rain per year through the compact soil.
0.80 inch of rain per year through the loose soil.

In a fourth experiment the same method as before was pursued, but the cylinders were shorter, the columns of soil having the depths of 1, 2, 4, and 6 inches. Two soils were used—Leonardtown loam with an initial water content of 1.0 per cent of moisture, and Podunk fine sandy loam with an initial water content of 0.4 per cent. Eight cylinders were used, the four for each soil having the various depths noted above. The experiment ran for one hundred and forty days, or from February 12 to July 2, 1906, i. e., during a period when the outside air was, on the whole, becoming moister and warmer.[a] The amounts of water taken up by the soil were much less than in the last experiment, and it has not been thought worth while to plot the results with the same detail. The results are summarized in Table II.

TABLE II.—*Loss of water by evaporation from below soil columns at different depths.*

Soil.	Depth in inches.	Initial porosity.	Loss from cups.	Loss from whole system.	Gain by soil.	Final water content.	Loss from cups in inches of rain per year.	K.[b]
			Grams.	*Grams.*	*Grams.*	*Per cent.*		
Leonardtown loam	1	0.54	120.3	119.4	0.9	1.6	2.71	4.88
Do	2	0.51	71.0	70.0	1.0	1.3	1.60	4.48
Do	4	0.49	42.2	38.0	4.2	1.7	0.95	4.57
Do	6	0.51	30.7	26.5	4.2	1.5	0.69	4.72
Mean		0.513						4.66
Podunk fine sandy loam	1	0.48	111.9	112.6	−0.6	0.1	2.52	4.54
Do	2	0.46	70.7	70.7	±0.0	0.4	1.59	4.45
Do	4	0.41	41.3	40.4	0.9	0.5	0.93	4.45
Do	6	0.46	29.7	28.4	1.3	0.5	0.67	4.56
Mean		0.453						4.50

[a] It is to be remembered that we are speaking here of the absolute and not the relative humidity of the air; for the partial pressure of the water vapor in the air, which is the main factor in regulating the diffusion, is determined by the absolute and not the relative humidity.

[b] K = (Depth of soil + 0.8 inch) × (loss per year).

It will be seen that the rates of loss from the cups below the two soils are in each case nearly the same for any given depth of soil, being slightly less for the Podunk fine sandy loam than for the Leonardtown loam.

It might at first sight be expected that the rate of loss should be inversely proportional to the depth of the soil through which the water vapor had to make its way in order to escape, or that the product of the rate of escape and the depth of the layer of soil should be constant. This, however, is by no means the case. The assumption involved is that the whole resistance encountered by the water vapor in its diffusion away from the cups is in the interstices of the soil, or, in other words, that the ventilation of the space immediately over the top surface of the soil is perfect. If we assume that the ventilation by the fan was imperfect, so that the water vapor after leaving the soil had still to encounter a certain amount of resistance to diffusion, and if we estimate this resistance as equivalent to that of a layer of the soil 0.8 inch thick we find satisfactory results. In other words, the rate of loss of water from the cups was, for each soil, inversely proportional to the depth of the soil plus 0.8 inch. This is shown by the approximate constancy of the numbers in the last column of Table II, which were obtained by multiplying the rates of loss of water from the cups, in inches per year, by the corresponding depths of the soil columns each increased by 0.8 inch.

The results of all the experiments described, in so far as they bear upon the original question of the actual amount of water lost from soils by direct evaporation from points below the surface, are collected in Table III.

TABLE III.—*Loss of water by evaporation from below columns of different soils.*

Soil.	Initial moisture condition.	Depth.	Duration of experiment.	Initial porosity.	Rate of loss per year.
		Inches.	*Days.*		*Inches.*
Coarse sand	Air dry	2	4	0.45	4.3
Takoma lawn soil	10 per cent	12	53	0.44	0.12
Do	do	12	53	0.50	0.15
Do	do	12	53	0.52	0.17
Cecil clay	Air dry	12	441	0.46	0.60
Do	do	.....	441	0.66	0.80
Leonardtown loam	do	1	140	0.54	2.71
Do	do	2	140	0.51	1.60
Do	do	4	140	0.49	0.95
Do	do	6	140	0.51	0.69
Podunk fine sandy loam	do	1	140	0.48	2.52
Do	do	2	140	0.46	1.59
Do	do	4	140	0.41	0.93
Do	do	6	140	0.46	0.67

CONCLUDING REMARKS.

The first of the experiments showed that the open space below the sand was nearly saturated with water vapor and that it made no material difference whether this space was fed with water vapor from a free

water surface or from damp soil. Hence we may assume that in the other experiments also the vapor space below the soil column was sensibly saturated. It could therefore not have had any more water vapor in it if it had been fed by evaporation from moist soil, no matter what the depth of the moist soil might have been below the level in question, and consequently there could in no case have been a greater tendency for the escape of water vapor from this depth. The figures given for the rates of loss of water from the cups represent, consequently, the maximum amounts of loss by direct evaporation which could have occurred from moist soils of any thickness below the depths named and under the given conditions of temperature and humidity.

The first, third, and fourth experiments are virtually experiments on the effectiveness of mulches of various dry soils of varying depths and compactness, the mulch being made perfect and the conditions simplified by the entire elimination of capillary motion into the mulch from below. The loss from below the coarse sand, at the rate of 4.3 inches per year through 2 inches depth, shows that the protecting action of this mulch was by no means perfect, and with a more loose, open mulch the protecting action, supposing always that the capillary flow into and up through the mulch is negligible, would be still less perfect. With the finer-grained soils of the fourth experiment the direct evaporation loss was less. But since the capillary action in these soils would, if they were in actual contact with the moist soil below, be considerable, we can not conclude that these mulches would necessarily be more effective than in the coarse sand.

The second experiment, while the most complicated, corresponds most nearly to practical conditions, and here we see that the direct evaporation loss from below a 12-inch layer of damp soil is quite insignificant, amounting at most to only 1 inch of rainfall in six years.

B.—THE DRYING OF SOILS UNDER ARID AND HUMID CONDITIONS.

When a moist soil dries by contact with the air above it, the loss of water takes place by evaporation close to the surface, the amount lost by direct evaporation from points several inches below the surface being, in general, negligible. As the surface soil dries out, a moisture gradient is established, and the dry surface soil draws up water from the moister region below by capillary action. If this capillary flow of water be prevented or lessened, as by the use of a mulch, the escape of water is decreased, because the evaporation has, on the whole, to take place from farther below the surface, so that it encounters greater resistance and is slower.

The flow of water in a soil which is not very wet has to take place through the thin films in which a part of the water is distributed over the soil grains. If the soil becomes very dry, it is to be expected that

these films will become thinner or break, the resistance to capillary flow increasing very much in either case. Hence it is to be expected that in very dry soils capillary flow will be slow.

Suppose a soil could be made so dry that no capillary flow at all would take place in it even when it was in contact with a moist soil. Such a layer of soil would act toward the moist soil below it as a protecting mulch, precisely as the layers of dry soil acted in the experiments described in the preceding paper. A soil can probably not be made so dry as to lose its power of capillary conduction of water entirely, but we may get an approximation to this limiting case.

Suppose that after a rain the soil surface be exposed to very arid conditions with a high surface temperature and a hot dry wind. The surface of the soil will lose water much faster than it can be brought up from below by capillary action, and if the arid conditions be kept up, a layer of dry soil will be formed on the surface, which may be so dry as to act in the manner suggested—i. e., as a protecting mulch. We shall thus have an initial period of very rapid evaporation followed by a period of slow evaporation taking place largely from below the surface.

If the same soil had been subject to less arid conditions, the initial loss would have been less. The capillary flow from below would have been sufficient to prevent the surface soil from becoming dry enough to act as a mulch; hence, though the initial rate of loss would be smaller, that rate would not fall off so rapidly. It might even happen that in the long run the soil under the arid conditions would actually lose less water than the same soil under humid conditions. Some experiments bearing on these points have been made in this laboratory, and they will now be described.

The soils were packed in metal cylinders 48 inches long and 2½ inches in diameter. The lower 42 inches of the soil were tamped down hard to imitate as far as possible the condition of a subsoil, while the top 6 inches were put in loosely in a condition of good tilth, the soil being moderately moist and in what would be good condition for growing a crop. Each cylinder had a side tube at the bottom through which water could be introduced. These tubes were kept stoppered except when they had to be opened to put in the water, which was done often enough to keep the bottom of the soil column always saturated with water. The conditions thus corresponded to those of a soil where the water table, or level of the ground water, was always at a distance of from 46 to 48 inches below the surface.

By means of an electric fan a current of air was blown over the top surface of the soils. To imitate arid conditions, one current of air was heated, without changing its absolute humidity, to a temperature about 50° to 60° F. above the room temperature. To imitate the high surface temperatures of soils under the strong sunshine of arid climates,

the top inch and a half of the cylinders under the hot air was heated, by heating coils surrounding them, to about the same temperature as the hot air. The breeze, of about 3 miles per hour, was kept going all the time. The heating current was turned on for six hours a day, except on Sundays and holidays.

In each experiment four cylinders were filled with the same soil in as nearly as possible the same manner. Two of these (humid conditions) were placed under the current of air at room temperature. The other two (arid conditions) were placed under the current of hot air and were provided with the heating coils. The cylinders were weighed every day except Sundays and holidays and a record kept of their losses of weight as well as of the weights of the water added at the bottom. In order to have some idea of the evaporation from water under the cold and the hot air, two water checks were run. Each of

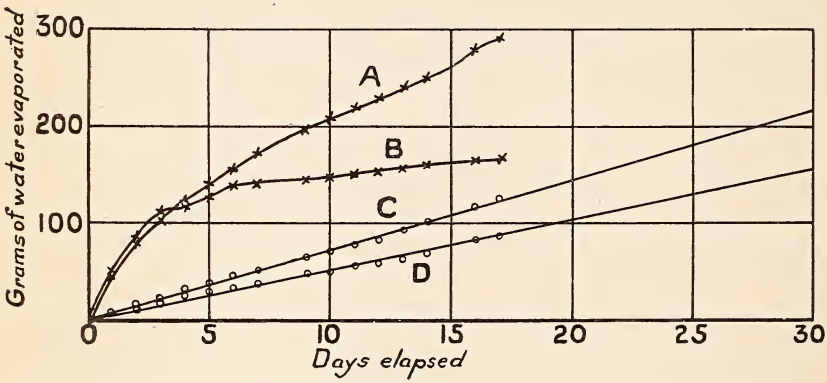

FIG. 3.—Evaporation of water from Takoma loam soil fed with tap water: A, soil under humid conditions; B, soil under arid conditions; C, water under arid conditions; D, water under humid conditions.

these consisted of a can 3 inches in diameter nearly filled with water and with its top covered with three layers of cheese cloth to diminish convection currents in the can. One of these cans was placed with each pair of cylinders, the relative positions being similar for each set. The cans were weighed at the same times as the soil cylinders and a record kept of their losses in weight.

In the first experiment the soil was a sandy loam from Takoma Park, Md., wet at the start with 18.6 per cent of water. The initial porosity of the lower 42 inches of soil, considered as dry (i. e., the fraction of the total space not actually occupied by solid soil grains), was 0.478. For the top 6 inches the corresponding value was 0.578. The cylinders were fed with tap water.

The results of this experiment, which ran for seventeen days, or from April 8 to April 25, 1905, are shown in figure 3, in which the abscissas represent the time elapsed in days since starting the experi-

ment, and the ordinates represent grams of water lost. The values plotted for the soils are the means from the two duplicate cylinders.

Upon examining this figure we see that the loss from the soil has in fact been somewhat greater at first under arid conditions than under humid conditions, but that this rate has so decreased that after about four days the total loss under the arid conditions becomes and remains less than under the humid conditions in spite of the fact that the evaporation from the water check was considerably higher under the hot than under the cold air. The rates of evaporation from the soils for the last ten days are 11.2 inches of rain per year under arid conditions and 51.6 inches of rain per year under humid conditions.

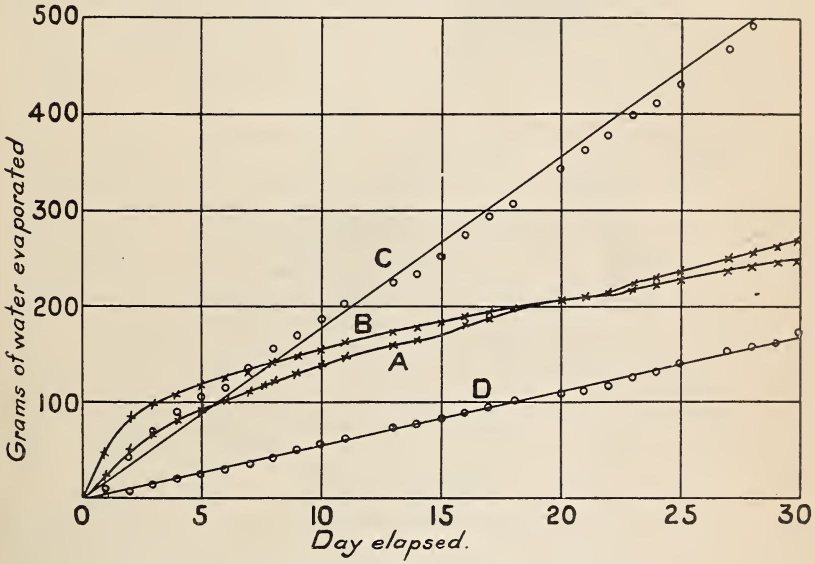

FIG. 4.—Takoma loam soil with alkali: A, soil under humid conditions; B, soil under arid conditions; C, soil under arid conditions; D, soil under humid conditions.

It thus appears that, under this imitation of arid and humid conditions, our a priori reasoning has been justified and that the arid conditions have actually established a dry surface layer which has to a considerable degree protected the soil below it from further loss.

In a second experiment, performed in the same way, the same soil was used, but was originally moistened, and thereafter fed at the bottom, with a N/10 NaCl solution. The initial percentage of moisture in the soil was 15.5 and the porosities were about the same as before. The experiment ran for sixty-six days, or from August 30 to November 4, 1905. The results for the first thirty days are shown in figure 4.

The results are qualitatively the same as before. The evaporation from the soil under the arid conditions is much more rapid at first,

but after about three days have elapsed the rate of loss is less under arid than under humid conditions, so that the total loss under the humid conditions gradually overtakes that under the arid conditions, passing it after about twenty-one days. The remaining observations, from thirty days on, showed no further changes in direction of any importance.

The fact that the crossing point of the curves, or the time at which the total evaporation under the humid conditions overtakes that under the arid conditions, comes so much later here than in the former experiment, when the soil was fed with water instead of salt solution, looks at first sight as though the salt had a considerable influence on the results. This conclusion, however, can not be drawn with any certainty, for we see from the water check curves that in the present case the evaporation from the water was 3.2 times as rapid under the hot

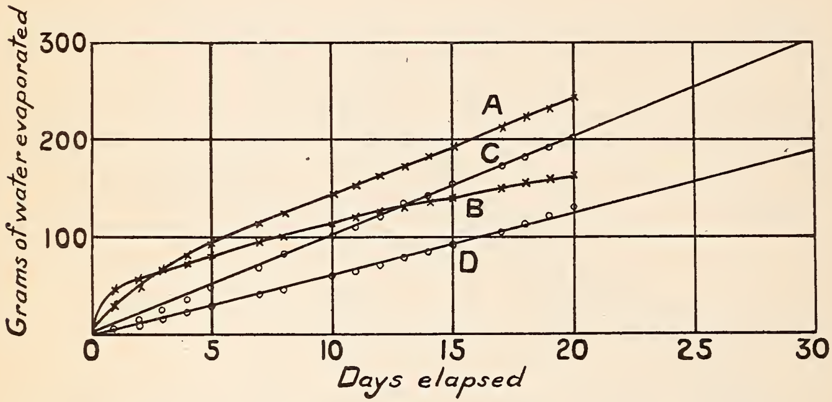

FIG. 5.—Evaporation of water from Podunk fine sandy loam with tap water: A, soil under humid conditions; B, soil under arid conditions; C, water under arid conditions; D, water under humid conditions.

as under the cold air, whereas in the previous experiment it was only 1.4 times as rapid, so that the two experiments are not comparable quantitatively. The rates of loss during the latter part of the experiment were 29.1 inches of rain per year under humid conditions and 20 inches under arid conditions.

In a third experiment the soil used was Podunk fine sandy loam fed with tap water, and wet at the start to a water content of 12 per cent. The experiment ran from November 24, 1905, to January 9, 1906, or forty-six days. The results for the first twenty days are shown in figure 5; they are qualitatively the same as before.. The evaporation is at first much more rapid under the arid conditions, but the rate soon falls off so that in about three days the total evaporation under humid conditions overtakes that under arid conditions and is thereafter increasingly greater. The mean rates of loss, during the later part of

the observations shown, are about 44 inches of rain per year under the humid conditions and 20 inches per year under the arid conditions, the loss from the water check having been 1.6 times as rapid under the hot as under the cold air.

A fourth experiment differed from the third only in the facts that a N 10 NaCl solution was used instead of tap water, and that the initial water content of the soil was somewhat higher, namely, 19.5 per cent. The experiment ran for fifty days, beginning April 25, 1905. The results up to thirty days are shown in figure 6, and are qualitatively like the previous ones. The crossing point of the curves comes later here, the evaporation under the humid conditions not overtaking that under arid conditions for forty-one days. The curves when plotted in full are entirely similar to those already given, and it has

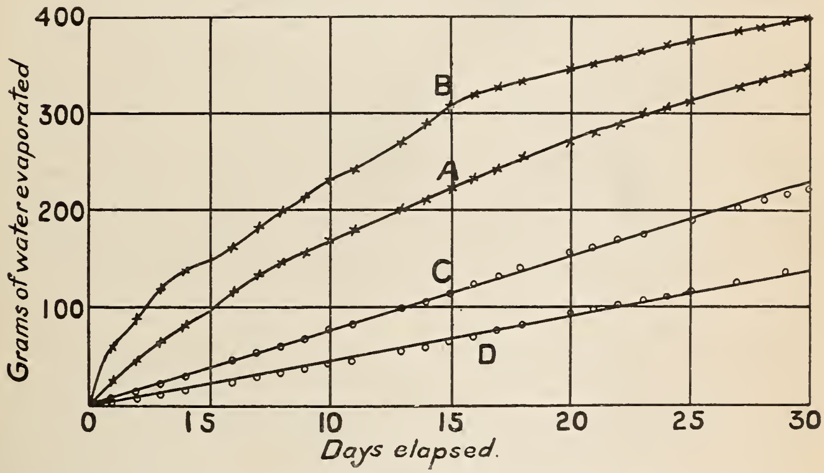

FIG. 6.—Evaporation of water from Podunk fine sandy loam with alkali: A, soil under humid conditions; B, soil under arid conditions; C, water under arid conditions, D, water under humid conditions.

not been thought necessary to plot the observations beyond thirty days. The rates of loss during the later part of the observations shown in figure 6 are 34 inches per year for the humid conditions and 26 inches for the arid conditions, the evaporation from the water check having been about 1.65 times as rapid under the hot as under the cold air.

In the case of these two experiments on the Podunk fine sandy loam we again find that with the salt solution the crossing point of the curves comes much later than with water—forty-one days instead of three days. The losses from the water checks bore in this case almost the same ratio for the two experiments, namely, 1.62 and 1.65, but the fact that in the third experiment the soil started with 12 per cent of water, while in the fourth it started with 19.5 per cent again makes

the experiments not strictly comparable, so that we can not say with certainty that the difference is due to using salt in one case and not in the other.

The ideas with which we started seem on the whole to have been well founded. It appears that under very arid conditions a soil automatically protects itself from drying by the formation of a natural mulch on the surface.

C.—CAPILLARY ACTION IN SOILS.

I. INTRODUCTION; NATURE OF THE PROBLEM.

§ 1. *Capillary water.*—If a soil be saturated with water and then allowed to drain while protected from evaporation, it will, after losing a certain amount of drainage water by percolation under the action of gravity, reach a steady state in which no further loss takes place, the remaining water being held in the soil by capillary action, partly in drops at the points of contact of the soil grains and partly in thin films on the surfaces of the grains. The amount of this "capillary water" retained by the soil, depends on various circumstances. It depends to a certain degree on the nature and amount of the substances which dissolve from the soil into the water; it also depends on the temperature. But aside from these two influences, which are secondary in our present considerations, it depends primarily on the depth of the soil to the level of free drainage or of standing ground water, on the texture or ultimate fine grainedness of the soil, and on its structure, i. e., its condition as regards granulation into compound particles and as regards arrangement or packing of these particles. The soil exerts a certain attraction sufficient to hold the water against the action of gravity which tends to drain it perfectly dry. This attraction must depend on the amount of water in the soil, for if there is more than a certain amount the excess drains away.

§ 2. *Capillary flow of water.*—If the surface of the soil be now exposed to evaporation, drying takes place, most at the surface but to some extent also at lower levels. This loss of water at lower levels is greater than can be accounted for by the assumption that the water evaporates directly into the pores of the soil and then escapes as vapor by diffusion; for it has been shown by experiment[a] that the loss by direct evaporation from depths of a few inches in a moist soil is very small. What actually happens is that as the surface soil dries out, its attraction for water increases so that it sucks up water from the moister soil below, with which it was previously in equilibrium.

This phenomenon of the capillary motion of water is, of course, not confined to the vertical but may occur in any direction from wet

[a] See Section A of this bulletin.

to dry portions of the soil. To get a clear idea of what takes place, we must think of it as a current of water through the soil; and as a measure of the strength of the capillary current at any point we may take the amount of water which passes in one unit of time through a unit area of an imaginary plane surface perpendicular to the direction of the motion. For example, we may measure the current strength or current density in centigrams per day per square centimeter.

§ 3. *Flow of water in tubes.*—If water be allowed to flow through a pipe from an elevated tank, the volume of water that comes out of the pipe per second depends on two things—the head or difference of pressure between the two ends of the pipe, and the ease with which the pipe conducts water, depending on the cross section, length, form, etc., of the pipe. We have thus two factors to consider—the cause, or the difference of pressure, and the facility of yielding to the cause, or the power of the pipe to conduct water. Both combined determine the effect, i. e., the flow of water.

If the pipe be very small in diameter compared with its length, as in the case of a long capillary tube, the relations are quite simple. The rate of flow of water is directly proportional to the product of the difference of pressure between the ends of the tube and a quantity which we may call the conductance of the tube and which itself is proportional to the fourth power of the diameter of the tube and inversely proportional to its length. We have, in other words, a relation of the form.

$$\text{Current} = (\text{Conductance} \times \text{Difference of pressure}) \times \text{Constant}.$$

§ 4. *Flow of heat; Fourier's law.*—If one end of a metal bar be placed in the fire, heat flows out along the bar, as is shown by the fact that parts of the bar at some distance from the fire become warmer, especially if loss of heat from the sides of the bar be diminished by wrapping it in a poor conductor such as asbestos. If all loss of heat from the sides of the bar be prevented, the current of heat is equal to the product of the difference of temperature of the ends of the bar (corresponding to the difference of pressure in the case of flow of water) and the thermal conductance of the bar, depending on its length, cross section, and composition.

The strength of the current at any given point, or the number of units of heat that pass in one second through any cross section of the bar, is equal to the temperature gradient or fall of temperature per unit length, multiplied by the conductance of the bar per unit length. If the bar have unit cross section the conductance per unit length is a property of the material, and is known as the thermal conductivity or specific thermal conductance of the material in question.

If we let K be this conductivity, i. e., the thermal conductance of a cube of the material 1 cm. on each edge, for a current of heat

flowing through it perpendicular to two opposite faces; and if we let H be the current density, i. e., the number of heat units which pass in one second through 1 sq. cm. perpendicular to the direction of the current, we have the equation

$$H = K G$$

in which G represents the temperature gradient at the given point in the given direction. This equation, which may also be written in the form

$$H_s = -K \frac{\partial T}{\partial S}$$

where T is the temperature and S the direction of the current, is the mathematical expression of Fourier's law, the fundamental principle on which the theory of heat conduction is based.

§ 5. *Constant electric currents; Ohm's law.*—If a steady electric current be flowing through a wire the strength of the current or the quantity of electricity that passes a given cross section of the wire per second is equal to the product of the difference of potential or electrical pressure between the ends of the wire and its electrical conductance. The difference of potential is the driving force or cause. The conductance of the wire (proportional to its cross section and inversely proportional to its length) represents the power of the wire to conduct a current or to yield to the driving force. The product of the two determines the effect produced—the current.

The current density at any point is equal to the potential gradient at that point—i. e., the fall of potential per unit distance parallel to the current, multiplied by the electrical conductivity or specific electrical conductance, which, as in the thermal problem, is defined as the conductance of a unit cube of the given material for a current perpendicular to two opposite faces. If we let L be the conductivity and C the current density, we have the equation

$$C = LE$$

where E is the potential gradient. This equation, which may also be written in the form

$$C_s = -L \frac{\partial V}{\partial S}$$

where V is the potential and S the direction of the current, is a mathematical expression of Ohm's law, which plays the same part in the theory of constant electric currents as Fourier's law in heat conduction. Since the two laws are identical in form, the mathematical treatment of problems in the two subjects is also identical.

§ 6. *Capillary currents through soils.*—If a column of dry soil, in a glass tube covered at the bottom with fine wire gauze, be placed with its lower end in a dish of water, the water begins to rise in the soil,

the dry soil sucking water away from the moister soil below. If the tube be not too long, and if its upper end be exposed to evaporation into the air, a steady state of flow is finally established in which the percentage of moisture at any level in the soil does not change with the time. When this state of affairs has been set up the amount of water that reaches any level from below must be the same as the amount that leaves that level to the soil above, and the same as evaporates from the surface at the top of the column. If the tube is of constant cross section, there is thus the same capillary current density, measured in grams per second through 1 sq. cm., at every point in the tube. Before this steady state is set up, the capillary current is not constant and the water content at any given point in the tube varies with the time.

If, after the soil at one end of the tube has been thoroughly wet, that end be covered and the tube laid horizontal, the capillary flow will continue, the current being from places of high to places of low water content, just as heat flows from high to low temperature, or electricity from high to low potential. In this case the current will not stay constant at any one point; the state of affairs will be variable and not steady.

§ 7. *Formulation of the problem.*—In studying this capillary conduction of water through a soil that is not uniformly moist, it is natural to be guided by the analogy of thermal and electrical conduction. The driving force or cause of the current is the different attraction for water of two portions of the soil that are not equally moist. The conductance of the soil is the facility with which it allows water to flow through it, or its power to conduct water.

Let Q be the capillary current density at any point—i. e., the mass of water which passes in one second through 1 sq. cm. of an imaginary surface perpendicular to the direction of flow. Let ψ be a quantity which measures the attraction of the soil at any given point for water. Then the gradient of attraction, which we may denote by S, is the amount by which ψ increases per centimeter in the direction of the current, by reason of the fact that the water content of the soil decreases in that direction. Let λ denote the capillary conductivity of the soil. Then we may write, in formal analogy with Fourier's and Ohm's laws,

$$Q = \lambda\ S$$

The analogy, however, is only formal. In the first place, the thermal and electrical conductivities of a given piece of material are independent of the strength of the current and, in general, only slightly dependent on the temperature and other outside circumstances, so that for most purposes they may be treated as constants. The capillary conductivity, however, we have every reason to expect to be

largely dependent on the water content of the soil, and therefore variable, not only from point to point in the soil, but also with the time at any given point. For it is not to be expected that the ease with which water flows through the soil will be independent of the extent and thickness of the water films through which—i. e. along which—it has to flow.

Furthermore, the other factor in the equation, namely, the gradient S, is not the space variation of a simple and directly measurably quantity like a head of water, an electrical potential, or a temperature. It is the gradient of a quantity ψ, the attraction of the soil for water; and ψ depends in some as yet unknown way, differing from soil to soil, on the water content of the soil, which can itself be measured only by tedious and not very accurate methods.

The general problem of investigating the laws of the capillary flow of water through soils is thus very much more complex than the corresponding thermal and electrical problems. The analogy of the Fourier-Ohm law helps us frequently in a qualitative way, and in some simple cases quantitatively also, but the attack on the complete solution of the capillary problem has been only begun.

It seems advisable, before attempting to describe and to analyze the experiments that have been made on capillary flow, to consider the quantities ψ and λ separately in the light of such data as are available. The quantity ψ, the measure of the attraction between soil and water, will be considered first.

II. THE CAPILLARY POTENTIAL.

§ 8. *Definition of the capillary potential.*—As we have already seen, the attraction of the soil for water depends on the water content, which we shall hereafter denote by θ. We can measure θ directly, if not very conveniently. We want, therefore, to find how ψ, the measure of the attraction, which we can not observe directly, is related to θ; i. e., we want, if possible, to get an idea of the form of the relation,

$$\psi = \psi\ (\theta)$$

If a perfectly dry soil be moistened, a rise of temperature may occur; there is in many cases a sensible development of heat during the wetting. If the soil be already slightly moist, the addition of more water produces no further heating or cooling, unless the soil contains abnormally large amounts of soluble salts. We shall make it a condition, in the reasoning which follows, that all the soils to be considered shall contain at least this small percentage of water beyond which further wetting produces no sensible thermal effects, so that the motion of water in them shall not of itself cause any development or absorption of heat, except that due to the dissipation of energy by the viscous resistances.

We shall assume that if we could, by purely mechanical means, pull a definite mass of water away from a definite mass of moist soil of a given moisture content, we should have to do a definite amount of mechanical work; and that if we then let the water and the soil come together again in obedience to their mutual attraction, we should, in principle at least, and if we could construct appropriate mechanism, be able to get back the same amount of work that we had to do in separating the water from the soil. This amounts to assuming that the attractive forces between the soil and the water are conservative, or that they have a potential. It is obvious that a rigorous treatment of the subject, with no restrictions imposed on either the water content or the soluble salt content of the soil, would have to use thermodynamic reasoning. But the simple conception of a mechanical potential will suffice for present purposes, though it is not impossible that with more comprehensive experimental data available we should have to use thermodynamic potential or the free energy.

Let θ be the water content of the soil measured in per cent of the dry weight, i. e., in centigrams of water per gram of dry soil. We might also measure the water content in grams per cubic centimeter. The percentage scale has been adopted, because the experimental data are most easily expressed in this way.

Let us consider a mass of moist soil containing 1 gram of solid matter and θ centigrams of water, or θ per cent. Let a small mass of water $\delta\theta$ be pulled away from the soil, so as to reduce its water content by $\delta\theta$.

Let $\psi\,\delta\theta$ be the work that we have to do to pull this water away. Then we may define ψ as the work required per centigram to pull water away from the mass of soil. We shall call ψ the capillary potential of the soil.

The value of ψ, for a given state of packing, temperature, etc., depends only on the water content, decreasing as θ increases. When the soil is completely saturated with water, its pores being full, water will begin to drain away from it at the first opportunity; it takes only an infinitesimal amount of work to remove a finite mass of water or $\psi=0$. If we let θ_0 be the water content of the soil when saturated, we have, therefore, the condition

$$\psi=0 \text{ when } \theta=\theta_0.$$

The capillary potential for a given water content varies from soil to soil; the retentiveness of different soils, or even of the same soil in different states of structure, is different. To put it in another way, if we subject the different soils to the same force, gravitational or other, tending to pull water away from them, we find that this force drains some soils dryer than others. But the final value of the capillary potential must be the same in all, because it just balances the same out-

side pull. Hence in some soils the water content has to be run down lower than in others to raise the capillary potential to a given value. These soils, in other words, are less retentive of water than the others.

§ 9. *Method of determining* ψ.—Consider a column of soil of uniform packing, with the water table at a constant level. Let there be no evaporation from the top surface, and let the soil and the water have been in contact so long that there is no further change in the distribution of water in the soil, i. e., let there be a state of equilibrium with no flow. Let x be the height of any given point above the water level where the soil is saturated. Then we have

$$\text{for } x=0,\ \theta=\theta_0 \text{ and } \psi=0 \quad . \quad . \quad . \quad . \quad . \quad . \qquad (1)$$

Let an infinitesimal mass of water δm, measured in centigrams, be moved by an outside impressed force from the level x to the level $(x+\delta x)$. The work done by the impressed force against the capillary attraction of the soil in removing the water from the soil at x, where the potential is ψ, is $\psi\delta m$. The work done by the capillary attraction against the impressed force as the water unites with the soil at the level $(x+\delta x)$, where the potential is $\left(\psi+\frac{\partial\psi}{\partial x}\delta x\right)$, is $\left(\psi+\frac{\partial\psi}{\partial x}\delta x\right)\delta m$. The total work done by the impressed force against the capillary forces is therefore

$$W_1=\psi\delta m-\left(\psi+\frac{\partial\psi}{\partial x}\delta x\right)\delta m$$

$$=-\frac{\partial\psi}{\partial x}\delta x\delta m.$$

At the same time the mass of water δm has been raised a distance δx against the force of gravity, which is $\left(\frac{g}{100}\delta m\right)$ dynes. Hence the work done by the impressed force against gravity is, in ergs,

$$W_2=\frac{g}{100}\delta m\delta x.$$

But by the principle of virtual work, if the system be in equilibrium, the total work of the impressed force during an infinitesimal change of the arrangement of the system must be zero. Hence we must have

$$W_1+W_2=0$$

or

$$-\frac{\partial\psi}{\partial x}\delta x\delta m+\frac{g}{100}\delta m\delta x=0$$

which when rearranged gives us

$$\frac{\partial\psi}{\partial\alpha}=\frac{g}{100}$$

or

$$\frac{\partial\psi}{\partial\alpha}=A \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \qquad (2)$$

If the units of measurement are changed the value of A will change also, but the relation will remain the same in form, A being a constant when once we have decided upon the units to be used.

What we want to find out is how the capillary potential depends on the wetness of the soil. Now the water content θ depends on the height x, so that the height may also be regarded as a function of the water content. Hence we have

$$\frac{\partial\psi}{\partial\theta}=\frac{\partial\psi}{\partial x}\cdot\frac{\partial x}{\partial\theta}$$

whence by equation (2) we get

$$\frac{\partial\psi}{\partial\theta}=A\frac{\partial x}{\partial\theta} \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (3)$$

By integrating this equation we get

$$\psi=Ax+B \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (4)$$

But by equation (1), $\psi=0$ when $x=0$, so that $B=0$, and we have simply

$$\psi=Ax \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (5)$$

Now, the relation of x and θ may be determined by analyzing the soil column. Hence we may plot the values of the height x as ordinate against the observed water content θ as abscissa. But by equation (5) the ordinates of this curve differ from the values of ψ for the same water contents only by the constant multiplier A, which is the same for all soils. Such curves may, therefore, be considered as giving the relative values of ψ since they give directly the values of $\frac{\psi}{A}$.

Regarding the general form of the curves we may note that as x increases θ decreases. Hence the curves cut the axis of θ at points which give the values of θ_0, the saturation value of θ, and rise toward the axis of x. We shall now give some examples of the form of these curves as found by experiment.

§ 10. *Forms of the* (ψ, θ) *curves.*—In figure 7 are shown the curves obtained from columns of six soils of various types, started simultaneously and analyzed after from fifty-three to sixty-eight days, when they had apparently ceased to take up any more water. The soils were in metal cylinders 48 inches long and $2\frac{1}{2}$ inches in diameter, closed at the top. Water was supplied through side tubes at the bottom, and the water level was, on the average, about $1\frac{1}{4}$ inches from the bottom of the tube, so that in plotting the curves the surface of the soil has been taken to be $46\frac{3}{4}$ inches above the water level. The curve for the Porters black loam is given in full on a smaller horizontal scale in figure 8.

The observations for the bottoms of the tubes, where the soils are very wet, are always of somewhat doubtful accuracy. It is probable also that the soils had not reached a state of equilibrium with the water and that the curves would have been somewhat different if the experiment had run for a year instead of for two months.

The soils were all somewhat moist at the start. The initial percentages of moisture and the "dry porosity," i. e., the fraction of the space not occupied by solid soil grains, are given in Table IV.

TABLE IV.

Type of soil.	Initial moisture.	Dry porosity.	Type of soil.	Initial moisture.	Dry porosity.
	Per cent.			*Per cent.*	
Podunk fine sandy loam	3.2	0.50	New Mexico dunesand	2.3	0.51
Porters black loam	19	0.67	Windsor sand subsoil	3.2	0.45
Cecil clay	10	0.52	Leonardtown loam	4.6	0.52

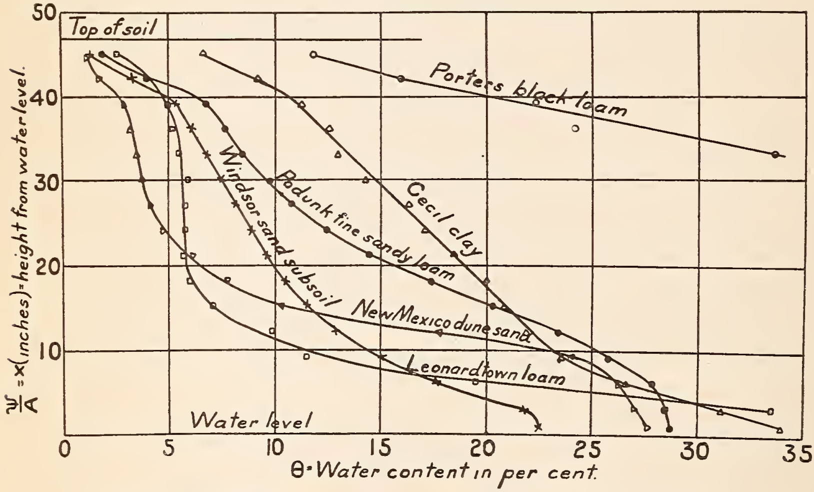

FIG. 7.—Distribution of water in 48-inch columns of soil after 53 to 68 days.

The values are only approximate, for in computing them it was assumed that the true density of the soil grains was the same for all, namely, 2.65. This assumption is doubtless pretty largely in error for the Porters black loam, a muck soil containing a great deal of organic matter.

In figure 9 are given the results of several experiments on Podunk fine sandy loam. Curve A is a repetition of the one already given in figure 7. Curves B, C, D, and E were obtained from experiments in which the tubes were open to evaporation at the top and a nearly steady state of flow upward had been established. In these experiments the top 6 inches of soil were packed more loosely than the lower 42 inches, so that in comparing the curves the upper parts should be

disregarded. The initial water contents and dry porosities were different in the different cases. Some of the important data are collected in Table V.

TABLE V.

Curve.	Initial water content.	Dry porosity.	Time.	Final rate of flow per year.	Remarks.
	Per cent.		*Days.*	*Inches.*	
A	3.2•	0.50	65	0	
B	12.0	0.35	46	44	Run simultaneously.
C	12.0	0.35	46	20	
D	19.5	0.34	50	35	Run simultaneously. Each the mean of two duplicates. N/10 NaCl used instead of tap water.
E	19.5	0.34	50	15	

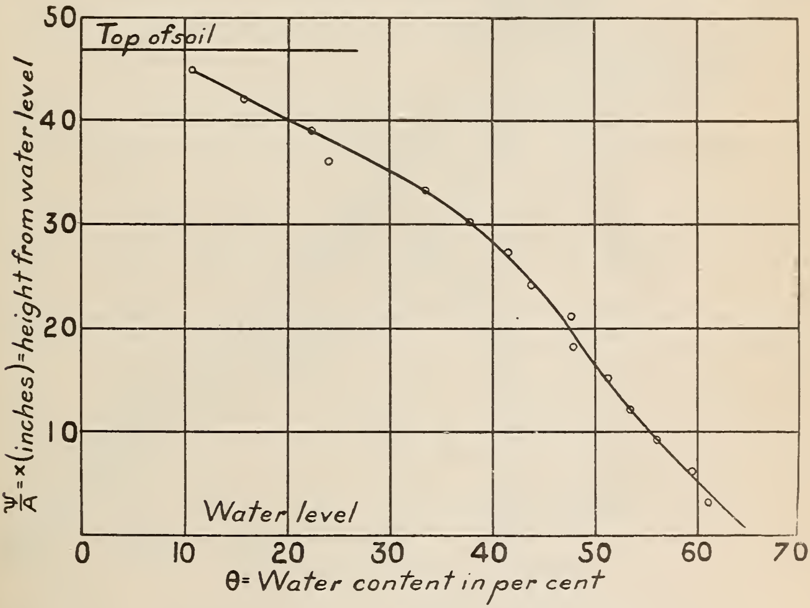

FIG. 8.—Distribution of water in Porters black loam.

It might appear that the curves B to E were not at all comparable with curve A because they were not obtained from experiments on equilibrium but from flow experiments. Now, in the experiments from which curves B and C were obtained, everything was precisely similar below a distance of 6 inches from the surface, except that for B the rate of flow of water (44 inches per year) was over twice as much as for C (20 inches per year). The curves, through the middle ranges at all events, are almost coincident. In the experiments from which the curves D and E were obtained there was the same similarity except that the flow in one case was somewhat more than twice as much as in the other (35 to 15). Here again the curves, while not quite so close together, are almost parallel. It appears, then, that when the rate of

flow is no larger than it was in these experiments changes in the rate of flow have hardly any influence on the final distribution of water in the soil when a steady state has been reached. Hence we naturally conclude, with great probability, that this final distribution is very nearly the same that would have been found if the tubes had been closed at the top and there had been no flow at all.

The curves are in fact quite similar in general shape. The difference between A and B or C is probably due to the different degrees of compactness of the soil. The more compact soil (curves B and C) holds its water more firmly, so that as we go up in the tube the water content θ does not fall off so fast as in the less compact soil (curve A). The difference between B and C on the one hand and D and E on the

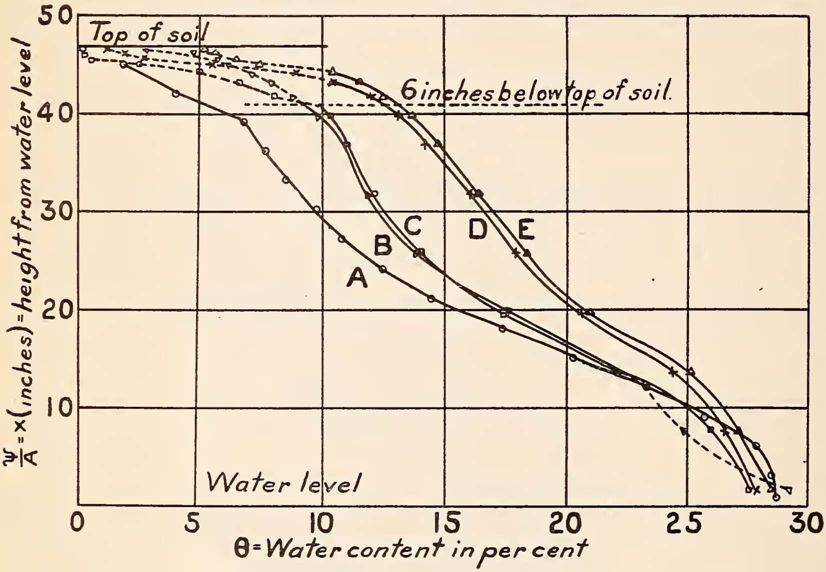

FIG. 9.—Distribution of water in Podunk fine sandy loam.

other can not be interpreted in this way, for the compactness was sensibly the same in all six tubes. The difference may be due to the fact that the initial water contents were different and that a steady state had not been reached, i. e., that the experiments were not run long enough. It may also be due to another cause, namely, the fact that in the experiment from which curves D and E were obtained the soil was wet with a N/10 NaCl solution instead of with water. It would not be surprising to find that the soil had a different attraction for such a solution and for water.

In figure 10 are shown the results from two experiments on Cecil clay. Curve A is a reproduction of the curve of figure 7. In the experiment from which curve B was obtained the cylinder was somewhat larger ($2\frac{2}{3}\frac{9}{2}$ inches internal diameter) and shorter. The tube

was not absolutely tight at the top, so that a slight loss was taking place. Data regarding the two experiments are given in Table VI.

TABLE VI.

Curve.	Initial water content.	Final mean water content.	Dry porosity.	Rate of loss per year.	Time.
	Per cent.	*Per cent.*		*Inches.*	*Days.*
A........	10.3	18.5	0.52	0	66
B........	18.3	30.4	0.59	1.9	328

The difference between these two curves is very great. It is too great to be attributed either to the small amount of flow taking place in one case or to the slight difference in compactness; in fact, both of these causes would be expected, a priori, to make the final results

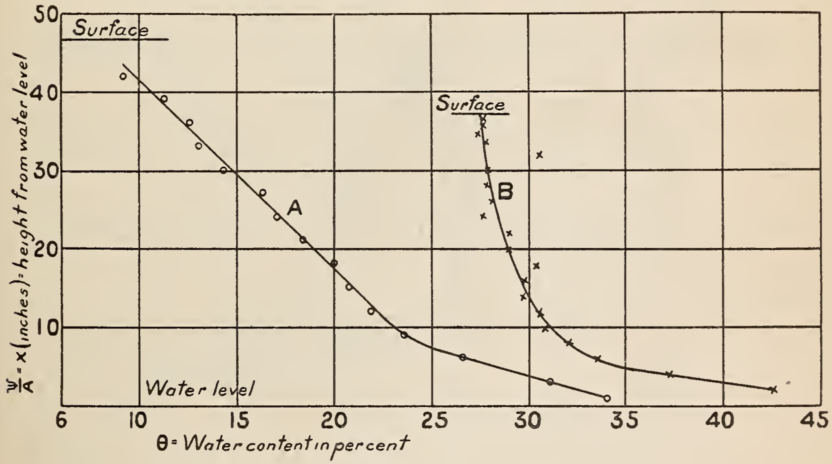

FIG. 10.—Distribution of water in Cecil clay.

differ in the direction opposite to that actually found. There must, therefore, be some other explanation. This is no doubt to be found in the difference of the duration of the experiments. In the sixty-six days of the first experiment, although the final rate of taking up water was so small as to appear negligible, the soil had not reached equilibrium with the water below it. Whether the three hundred and twenty-eight days of the second experiment were enough could be determined only by a repetition of the experiment, several tubes being run in duplicate and analyzed after various elapsed times. The irregularity of the observations shows the difficulty of working with a soil as heavy as this when the water content is high.

In figure 11 are shown two curves (A and B) for Leonardtown loam and two (C and D) for Takoma lawn soil, a sandy loam soil. The data for these experiments are given in Table VII.

TABLE VII.

Curve.	Initial water content.	Final mean water content.	Dry porosity.	Final rate of flow per year.	Time.	Remarks.
	Per cent.	*Per cent.*		*Inches.*	*Days.*	
A	4.6	10.3	0.52	0.0	63	Leonardtown loam, top 6 inches loose.
B	16.4	20.9	0.50	2.4	326	Do.
C	15.5	10.4	0.44	10.6	67	Takoma lawn soil fed with N/10 NaCl.
D	15.5	11.6	0.44	13.2	67	Do.

In comparing the two curves for Leonardtown loam (A and B) we find quite as great a difference as between the two curves of figure 10 for Cecil clay. The rate of flow was negligible in both cases. The difference in compactness was also very small. The difference is prob-

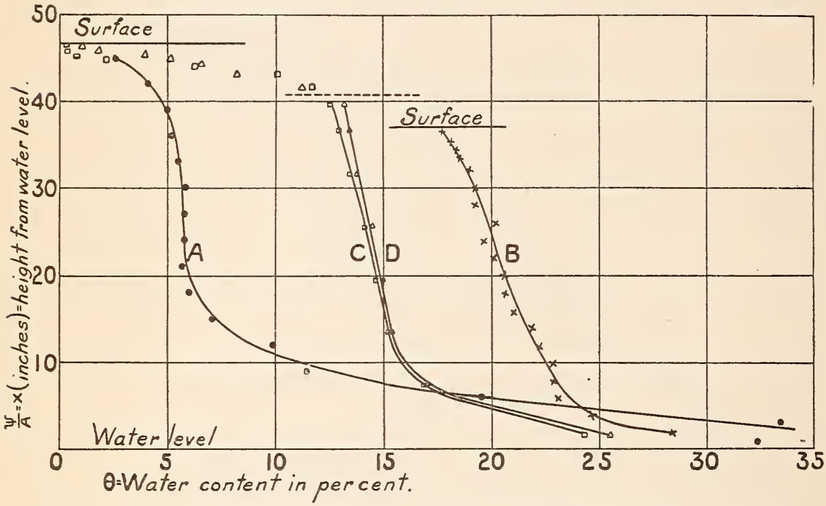

FIG. 11.—Distribution of water in Leonardtown loam (A and B) and Takoma soil (C and D).

ably due mainly to the fact that in sixty-three days the soil had not come nearly into equilibrium with the water, although its rate of taking up water had become so slow as to appear negligible at the time when the column of soil was taken down for the hygrometric analysis. Whether the three hundred and twenty-six days' duration of the experiment from which curve B was obtained were sufficient, could be determined only by further experiment.

The curves for the Takoma lawn soil (C and D) are quite similar in general form to those for Leonardtown loam. Each point plotted represents the mean obtained from measurements on two duplicate cylinders. All four were run simultaneously and the conditions below 6 inches from the surface were sensibly the same except that the final rates of flow, found for the last nineteen out of the total sixty-seven days were somewhat different. These rates of flow had apparently reached a nearly steady state.

As was the case with Podunk fine sandy loam we find that the curves do not differ much. We should expect that the more rapid flow would be accompanied by a more rapid rate of increase of water content with the depth. The opposite, however, is the case, curve C corresponding to a smaller rate of flow than curve D (10.6 to 13.2). We can therefore not attribute the difference, which is indeed not very pronounced, to the different rate of flow which thus again, when of the present small magnitude, appears to be of no importance in determining the shape of the curves.

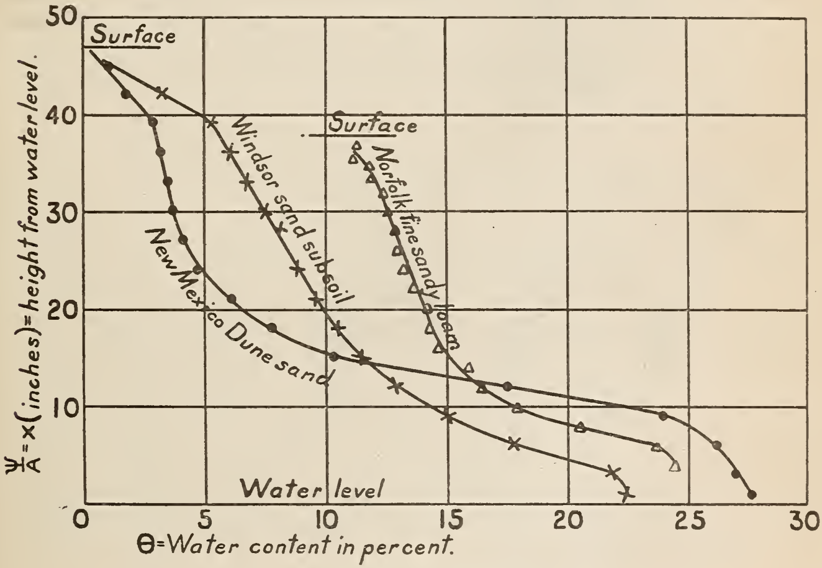

FIG. 12.—Distribution of water in sandy soil.

In figure 12 are reproduced the curves of figure 3 for New Mexico dunesand and Windsor sand subsoil (curves A and B) and, in addition, a curve found in another experiment for Norfolk fine sandy loam. Data for these soils are given in Table VIII.

TABLE VIII.

Curve.	Initial water content.	Final mean water content.	Dry porosity.	Final rate of flow per year.	Time.	Remarks.
	Per cent.	*Per cent.*		*Inches.*	*Days.*	
A	2.3	10.7	0.52	0	65	New Mexico dunesand.
B	3.2	10.6	0.51	0	61	Windsor sand subsoil.
C	10.6	15.3	0.44	1.0	324	Norfolk fine sandy loam.

The conditions for A and B are almost identical, and the pronounced difference in shape indicates a real difference in the behavior of the soils toward water. It would probably persist if the experiment had

been continued longer. Judging from what has already been said it is highly probable that if the time had been longer the curves would have been shifted to the right.

Curve C, for which the time was three hundred and twenty-four days, probably represents pretty nearly the final equilibrium distribution in the Norfolk fine sandy loam.

§ 11. *Remarks on the forms of the curves for different soil types.*—It appears probable, on the whole, that in all the experiments mentioned, where the duration was not more than about two months, the final state of equilibrium had by no means been reached, and that such experiments should be run much longer. The results of the one experiment that ran for a much longer time, about eleven months, are again reproduced, for the sake of comparison, in figure 13, the data being given in Table IX.

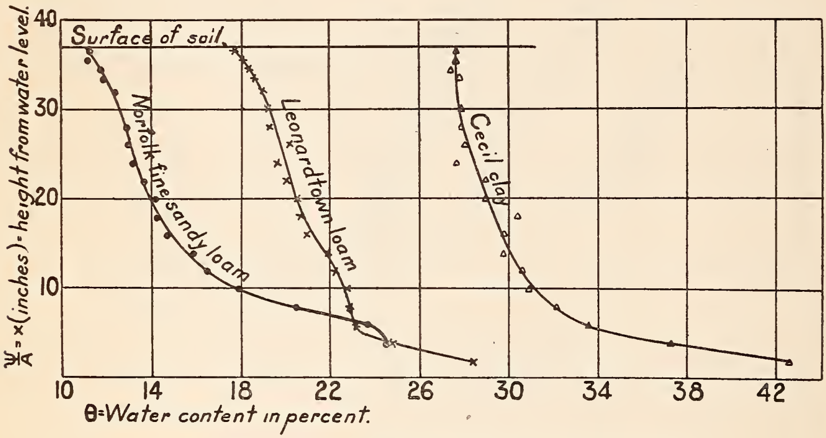

Fig. 13.—Distribution of water in soils after ten and one-half months.

Table IX.

Soil.	Initial water content.	Final mean water content.	Dry porosity.	Final rate of flow per year.	Time.
	Per cent.	*Per cent.*		*Inches.*	*Days.*
Norfolk fine sandy loam	10.6	15.3	0.44	1.0	324
Leonardtown loam	16.4	20.9	0.50	2.4	326
Cecil clay	18.3	30.4	0.59	1.9	328

It seems improbable that these three curves would have changed very much if the experiment had run longer, though we can not be certain that they would not. Assuming, however, that the curves have approximately their final forms and that these forms are not sensibly affected by the slight flow that was going on, it is instructive to compare them.

In the first place, starting at the lower end of the curve, we have at first a part where the slope is small, i. e., where the water content

decreases rapidly with increasing distance above the standing water. The slope of this part of the curve is not very different for all these soils. This part of the curve merges gradually into a much steeper part and the slope of this steeper part is also about the same for all three. The most striking difference between the three curves is that the heaviest soil, Cecil clay, holds by far the most water, and the lightest, the Norfolk fine sandy loam, the least. This is of course precisely what we should expect, the water-holding power against a certain pull, gravitational in this case, increasing with the amount of very fine particles, and therefore of very fine capillary spaces in the soil.

Since the final water content of each soil even in the top layer is greater than the initial, it is evident that a greater duration of the experiment could only be expected to displace the curves toward the right, if it changed them at all. Hence in order that the curves in their final equilibrium form should reach the water content of the air-dry soils it is obvious that the diagram would have to be extended upward, probably a long distance. In other words, if such an experiment is to give us quantitative information on the value of the capillary potential ψ as a function of the water content θ down to low values of θ, the tubes must be very much longer than those used in this experiment. They should also of course be more carefully sealed at the top, and care should be taken to make sure that the duration of the experiment is long enough. The tubes ought also to be kept at a constant temperature.

§ 12. *Modification of the experimental method.*—In all the foregoing experiments the soil used was kept saturated at the bottom; in other words, the general equation (4) of section 9, namely,

$$\psi = Ax + B \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (4)$$

was subject to the condition that

$$\psi = 0 \quad \text{when } x = 0$$

so that it reduced to the simpler form

$$\psi = Ax \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (5)$$

If the soil had not been subject to these conditions we should have had to use the general equation, which may be put in the form

$$\frac{\psi - B}{A} = x$$

or

$$\frac{\psi}{A} - C = x \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (6)$$

where $C = \frac{B}{A}$ = constant.

This shows us that we may find the variations of ψ, or the values of ψ minus a constant, by observations on the final state of equilibrium-distribution of water in a column of soil with any given water content and closed at both ends. By analyzing the soil and finding the value of the water content at each height x, measured from the bottom of the tube, we can find those relative variations. The value of B may be found if we have a series of overlapping experiments, the last of which has the soil saturated at the bottom. It is thus not necessary to use one very long column of soil; a number of shorter ones may be made to serve the same purpose, and besides the greater convenience, the condition of constancy of temperature may be satisfied more easily.

III. THE CAPILLARY CONDUCTIVITY, λ.

§ 13. *General considerations; capillary and film water.*—Having now shown how it is possible to investigate the capillary potential ψ experimentally, we have to consider the capillary conductivity λ.

When a soil is moist but not very wet, part of the water is spread out in thin films on the surfaces of the grains and the rest is collected into drops around the points where the grains are in contact or very close to one another. For convenience we shall speak of this water in mass as "capillary water," to distinguish it from the "film water."

The interstices of the soil are filled with air, and, except in unusual circumstances or unless the soil be very wet, this air is everywhere in free communication with the outside air and is at ordinary barometric pressure. Each film of water thus has air of atmospheric pressure acting on one side of it, tending to squeeze it out thinner against the solid soil grain on the other side of the film. The surfaces of the drops of capillary water are also subject to the same pressure; but since their surfaces are everywhere concave outward the pressure in the capillary water is less than atmospheric.[a] Now the films are continuous with the drops of capillary water. Hence the result of the atmospheric pressure is to force water out of the films into the drops, decreasing the total mass of film water and increasing the mass of capillary water by the same amount.

This process, if it kept on, would finally squeeze the films out of existence: they would break and the water would be all in drops. We know, however, that this does not take place; the soil grains remain wet on their surfaces. When the films get down to the small thicknesses commonly designated by such terms as the "radius of molecular action," new forces come into play. There is an attraction or adhesion between the water and the solid soil grains which acts sensibly only through very minute distances but then increases rapidly as the distance decreases. With a given amount of water in the soil, the films will be reduced to a certain minimum thickness and no far-

[a] Bul. No. 10, Bureau of Soils, U. S. Dept. of Agr.

ther. If the total amount of water in the soil be decreased, the capillary drops will be smaller and the curvatures of their surfaces sharper. Hence the negative pressure in them will be greater and there will be more tendency for the water in the films to be forced into the drops, the films becoming thinner. But since we have reason to believe that the adhesive forces increase very rapidly as the distance decreases, the thickness of the films will not vary much with the total water content until the negative pressure in the capillary drops becomes very great, i. e., until the drops are very small and the soil very dry. Then we must expect the films to become much thinner or to break altogether; but for medium water contents it seems probable that the thickness of the films on the soil grains, in a state of equilibrium, does not vary very much. The proviso that there shall be a state of equilibrium is necessary, because the motion of the water along through the films toward the drops takes time, so that in a soil which has only recently been wet the films may be considerably thicker than they are later, even though no evaporation takes place. A soil freshly moistened may therefore have somewhat different physical properties from one which has been moistened for some time, even aside from changes due to solubility, change of solid phase, etc. It is possible that the changes observed in the electrical conductivity of a freshly moistened soil may be in part due to this gradual change of the distribution of water.

§ 14. *Change of conductivity with water content.*—When a soil is very wet the capillary water in it is continuous; the drops merge into one another at their edges. It is thus possible, starting on one side of a layer of the soil, to describe continuous paths which shall pass through the soil to the other side without having anywhere to pass out from the water-in-mass of the capillary water and through the films. A current of water could therefore flow through the soil existing always as water-in-mass and without having to flow through surface films on the grains. The cross section of any such tortuous channel through the soil will vary greatly from point to point, being in some places very small. We know, as already mentioned in section 3, that the conductance of a capillary tube for water depends on the fourth power of its diameter. Hence the resistance such a channel opposes to a current of water will be due almost entirely to the more constricted parts; or, in other words, the conductivity of the soil in this state of wetness, for a current of water through it, will be determined almost entirely by the fineness of the smallest spaces between the grains through which the water has to flow.

If the water content of the soil be gradually reduced the water retires more and more into these very fine spaces, which remain full. Hence if the total number of such capillary channels were not diminished the capillary conductivity would not change much. In fact, however, as the water content is reduced the number of continuous

paths through capillary water diminishes till finally most or all of the capillary drops cease to be continuous, and are, from this point onward, in communication with one another only through the film water which stretches from one drop to another over the intermediate surfaces of the solid soil grains.

Hereafter the flow of water has to take place for a part of its course, at all events, through the films. These are very thin—of the same order of thickness as soap-bubble films. One face of each film is against a solid surface, and the particles of water in actual contact with the solid may be regarded as fixed to the solid; for experiments on flow through capillary tubes have shown that the amount of slip under such circumstances is negligible if not rigorously zero. The other face of the film is in contact with the air. The particles of water forming this surface may be freely movable; but it seems more probable, from the general behavior of dirty surface films on water, that they also act as if fixed. In the second case we have water flowing in a very thin layer between two fixed surfaces to which it adheres: In the first case the thin layer of water is fixed by adhesion at only one face. In either case the viscosity of the water opposes the flow-and the resistance increases as the layer of water grows thinner.

We have seen that the thickness of the films probably does not change much until the soil gets so dry that the films begin to break. If we assume as an approximation that their thickness is constant, then for a given general arrangement of films the conductivity of the soil will be directly proportional to the mean width of film and inversely proportional to the mean length along the film traversed by the water. As the water content decreases, the distance from one capillary drop to another increases, i. e., the conductivity decreases. At first, just after the drops have ceased to afford continuous paths through capillary water, the distances through the films will be short, and any given abstraction of water from the soil, causing a decrease in the average size of the drops, causes an increase in the film length, which is large in proportion to the length already existing. As the drops get smaller and the films longer, a given reduction of the water content causes a smaller fractional increase in the length of film from one drop to the next. Moreover, since the films are very thin, when the average length of the path through them has become considerable, the resistance they oppose to the flow of water will be large compared with the resistance in the parts of the path which lie in capillary water, so that the total resistance, and therefore the total capillary conductivity of the soil, will be determined almost entirely by the films and only to a minor degree by the shape and size of the capillary drops.

If, finally, the water content of the soil be reduced still further till the soil approaches a state of complete dryness, the indications from

experiment are that the films either break or become so thin as to lose their conducting properties. The conductivity of the soil for a current of water through it will then decrease rapidly toward zero.

What then must we expect, in a general way, to be the course of the changes in the capillary conductivity λ as the water content of the soil θ is gradually decreased to zero from an initial value θ_0 when the soil is saturated? According to the general reasoning just given we must expect:

(1) Large conductivity at the start decreasing as the number of continuous paths through capillary water decreases.

(2) When most of these paths have been broken and the film paths are becoming of importance though still short, on the whole, a rather rapid decrease of the conductivity.

(3) As the film paths get longer a less rapid decrease of conductivity until,

(4) As the soil approaches complete dryness, the films begin to break or to lose the properties of liquid water, when the conductivity will fall rapidly toward zero.

The general nature of the relation of the capillary conductivity λ to the water content θ will, if the foregoing reasoning is correct, be such as is represented in the curve of figure 14. The point A of the curve shows the value of λ when the soil is saturated with water. As the water content decreases the capillary paths become fewer, the conductivity decreasing. At B the much smaller conductivity due to the incipient formation of film paths begins to be of importance and as the film paths become longer the conductivity falls rapidly toward the point C. From C to D the flow is taking place almost exclusively through film paths, but the length of these is not increasing so rapidly in proportion as before, so that the curve does not fall so rapidly from C to D as from B to C and is concave upward. At D the soil has reached such a state of dryness that the films begin to break and the conductivity falls toward zero, possibly reaching this value before the water content has become zero. The curve may, in other words, not go from D to the point O but to some point, F, to the right of O.

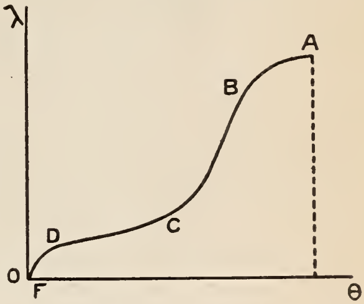

FIG. 14.—Variation of conductivity (λ) with water content (θ).

We have no data from which to plot the curve except in this qualitative way, so that the proportions may be quantitatively very far from right. It may be, for instance, that the part C D is very nearly flat and horizontal and that it occupies by far the greater part of the

whole length of the figure; or it may be that C D is so short that B C merges directly into D F. The relative lengths of various parts of the curve and their slopes may very probably vary greatly from soil to soil. We may say in general, however, that the part B C D is that of most importance to us in the study of the capillary motion of water through soils. For the part A B corresponds nearly to saturated soils, and the part D F to very dry soils. Experimental data are very hard to get in either of these regions, and, furthermore, agriculture is interested mainly in soils of medium water content, since ordinary crops can not grow under extreme conditions of wetness or dryness. We shall therefore direct out attention mainly to the section of curve B C D.

§ 15. *Special hypotheses regarding the relation* $\lambda=f(\theta)$.—As a check on the general reasoning of the last section, it is well to start from the beginning with some plausible assumption and see if we can work out mathematically a relation between the capillary conductivity and the water content, i. e., find a form for the equation

$$\lambda=f(\theta)$$

resulting from the hypothesis. For the reasons already stated, we shall disregard very high water contents and shall consider the soil to be in an ideal state where the continuity of the paths through capillary water has just ceased. We shall also make the simplifying assumption that under these circumstances, where any path along which water may flow through the soil lies partly through capillary drops and partly along films, the resistance to flow in the films is great in comparison with that opposed by the drops, so that the conductivity of the soil is determined almost entirely by the distribution, average length, etc., of the films, and only slightly influenced by the drops.

The capillary water held about the points of contact or of nearest approach of the soil grains is contained in spaces of various irregular shapes. These, however, may be assimilated to the following three general types:

(A) Ring-shaped wedges, where two soil grains touch at a single point. The simplest case would be the contact of two spheres of the same radius, or of a sphere with a plane.

(B) Prismatic or cylindrical wedges, where two soil grains are in contact or very nearly in contact along a straight line of finite length.

(C) Spaces of approximately conical or pyramidal form, such as may occur where three or more soil grains come together.

A close approximation to the true properties of a soil in its mechanical relations to water, both as regards its conductivity and as regards its water-holding power, or the capillary potential, might probably be obtained by supposing a certain fraction of the spaces in which the capillary water is held to have some one simple form of each

of the three types just mentioned. In the absence of sufficient experimental data we shall consider only one or two very simple cases.

§ 16. *Capillary water held in prismatic wedges.*—As a first and simplest assumption, suppose that the water behaves as if held in prismatic wedges, i. e., let us take the simplest example of type (B) and see what conclusions we can draw.

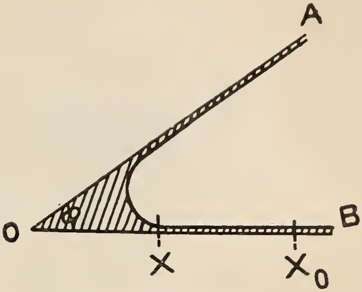

FIG. 15.—Capillary water held in a prismatic wedge.

Let $\overline{AOB}$ be an end view of one of the wedges, $\overline{OA}$ and $\overline{OB}$ representing edge views of the grain surfaces which, for simplicity, are represented as plane.

Let x be the extension of the capillary water from the vertex of the wedge. Let x_0 be the maximum extension, i. e., the extension if the volume of water were so increased that the capillary water just reached that from the next wedge to the right of B. Then the length of film belonging to this particular wedge $\overline{AOB}$ is (x_0-x).

Assuming the film thickness to be constant, and assuming that the resistance of the soil is sensibly all due to film resistance, we have for the conductance of such a film in the direction $\overline{OB}$

$$\lambda=\frac{\alpha}{x_0-x} \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (7)$$

where α is a constant depending on various circumstances, and, among others, proportional to the width of the film, i. e., to the length of the wedge in a direction perpendicular to the plane $\overline{AOB}$.

Neglecting the mass of water in the film in comparison with that of the capillary water, we also have for the volume of water present

$$\theta=\beta^2 x^2 \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (8)$$

where β^2 is a constant depending, among other things, on the angle φ of the wedge. If θ_1 represents the mass of water in the wedge when the soil is so wet that the film has just shrunk up to nothing, we have

$$\theta_1=\beta^2 x_0^2 \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (9)$$

Combining equations (7), (8), and (9), we get

$$\lambda=\frac{\alpha\,\beta}{\theta_1^{\frac{1}{2}}-\theta^{\frac{1}{2}}} \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (10)$$

If we suppose the water to be uniformly distributed through the soil in wedges of this sort, and suppose the flow to take place perpendicular to their edges, we may take λ and θ in equation (10) to be proportional to the capillary conductivity and the water content of the

soil as a whole, and equation (10) will still hold, only the value of the constant $\alpha\beta$ being changed by the new interpretation.

By differentiating equation (10) we find that

$$\left.\begin{aligned} &\text{when } \theta=0;\ \lambda=\frac{\alpha\beta}{\theta_1^{\frac{1}{2}}},\quad \frac{\partial\lambda}{\partial\theta}=\infty,\ \frac{\partial^2\lambda}{\partial\theta^2}<0 \\ &\text{when } \theta=\theta_1;\ \lambda=\infty,\quad \frac{\partial\lambda}{\partial\theta}=\infty,\ \frac{\partial^2\lambda}{\partial\theta^2}>0 \end{aligned}\right\} \quad . \quad . \quad . \quad . \quad . \quad (11)$$

The curve is tangent to the axis of θ at the point $\lambda=\alpha\beta\ \theta_1^{\frac{1}{2}}$ and rises with its convexity upward. It approaches the vertical line $\theta=\theta_1$, asymptotically running up to $+\infty$; i. e., concave upward. The first of these statements involves a physical absurdity, namely, a finite conductivity $\alpha\beta\ \theta_1^{\frac{1}{2}}$ for an absolutely dry soil. This is due to our having

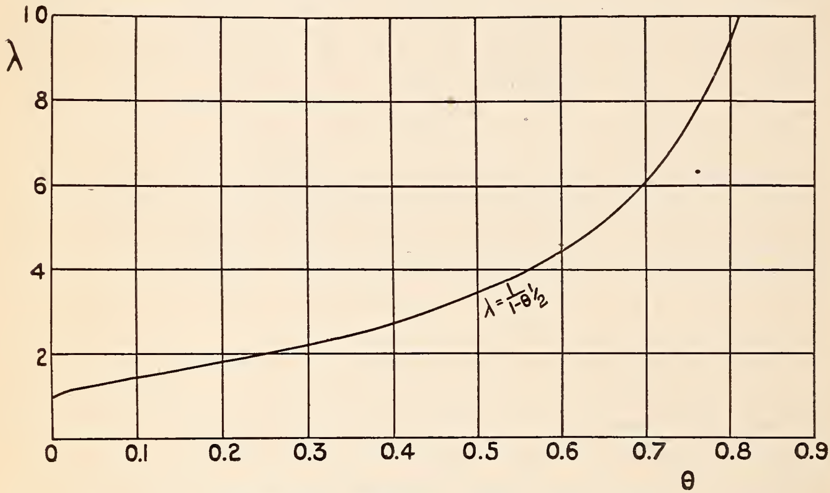

FIG. 16.—Relation of conductance (λ) and water content (θ).

neglected the volume of the film water altogether, which is, of course, not permissible when the total water content approaches zero. The absurdity might be avoided by introducing the appropriate correcting terms into the equations.

To plot the curve given by equation (10) we have to assign values to the constants $\alpha\beta$ and θ_1. To get a general idea of the form it is sufficient to let $\alpha\beta=1$ and $\theta_1=1$, so that we have

$$\theta=\left(1-\frac{1}{\lambda}\right)^2 \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (12)$$

In figure 16 a part of this curve is plotted. It will be seen that it has the same general form as the portion $\overline{BCD}$ of figure 15. The difference near the origin is due to the simplifying assumptions made, in which we have virtually refrained from considering the real state of affairs in a very dry soil.

§ 17. *Further considerations on film conduction.*—We may start from an altogether different point of view. Let us look upon the soil as an aggregate of spherical particles of equal sizes. When the equilibrium distribution has been reached, supposing gravity to be of negligible importance, the water will be collected in a number of equal drops about the points of contact of the spheres, connected by spherical films all of the same thickness. The problem is then mathematically the same as that of the flow of electric current in a spherical conducting sheet of uniform conductivity between a number of circular electrodes with their centers at fixed points on the sheet and their radii all equal but variable. The general solution of this problem would require much more mathematical work than it is worth while to go through when our results are to be at best only qualitative; but by simplifying assumptions we may help to clear our ideas.

As an approximation, consider the case of conduction in an infinite plane sheet between two circular electrodes of radius x with their centers at a distance $2x_0$ apart. (See figure 17.) This problem also we shall not solve completely. We shall merely consider what happens when the electrodes are very small or very large—i. e., when the capillary drops are so large as to be almost continuous or so small as to be on the point of vanishing altogether. This will give us results comparable with those expressed in equations (11) of section 16.

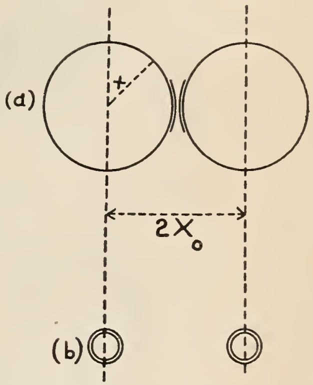

Fig. 17.—Diagram illustrating conduction between circular electrodes in an infinite plane sheet.

Consider first the case where the radius of the drops is nearly equal to x_0 its maximum value (figure 17, *a*). As x approaches x_0, the conductance of the film becomes dependent solely on the narrow strip of film immediately between the drops. If this could be treated as of constant length and uniform width, its conductance would be inversely proportional to its width—i. e., to $(x_0 - x)$. In fact, however, the length, too, is increasing, and since the conductance of such a strip is directly proportional to its length we shall have the conductance approaching the form

$$\lambda = \frac{Af(x)}{(x_0 - x)} \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (13)$$

Where $f(x_0)$ is finite and $f'(x) > 0$. The obvious conclusion is that if

$$x = x_0,\ \lambda = +\infty,\ \text{and}\ \frac{\partial \lambda}{\partial x} = +\infty \quad . \quad . \quad . \quad . \quad . \quad . \quad (14)$$

In the second extreme case, where x approaches zero, the conductance is determined more and more exclusively by the circumference of the drop or electrode, and it may be shown that if

$$x=0,\ \lambda=0 \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (15)$$

The consideration of $\frac{\partial\lambda}{\partial x}$ leads us to indeterminate forms which are difficult to handle, and as this end of the curve is of less importance and our hypotheses are farther from the truth, we shall content ourselves with the information contained in equation (15).

To pass from the foregoing case of conduction between two circular electrodes in a plain sheet to conduction in an aggregate of spherical sheets, we may remark in the first place that the conductivity of the aggregate will have in a general way the same variations as those already considered so that we may identify the λ of equations (13), (14), and (15) with the capillary conductivity of the soil. As regards the water content θ, it may be expressed in the form of a convergent series

$$\theta=\alpha_0+\alpha_1 x+\alpha_2 x^2+\text{ etc.} \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (16)$$

If $x=0$, $\theta=\alpha_0$, or α_0 is the amount of water contained in the films when the drops have shrunk up to nothing.

If $x=x_0$, $\theta=\theta_1$ the water content when the drops are about to run together. Furthermore, we have

$$\frac{\partial\theta}{\partial x}=\alpha_1+2\alpha_2 x+3\alpha_3 x^2+\text{ etc.} \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (17)$$

and if $x=x_0$, i. e., $\theta=\theta_1$

$$\frac{\partial\theta}{\partial x}=\alpha_1+2\alpha_2 x_0+3\alpha_3 x_0{}^2+\text{ etc.}=\text{constant.} \quad . \quad . \quad . \quad (18)$$

Hence, returning to equations (14) we see that if

$$\theta=\theta_1, \qquad \lambda=+\infty, \qquad \text{and } \frac{\partial\lambda}{\partial\theta}=+\infty \quad . \quad . \quad . \quad (19)$$

while from (15) if

$$\theta=\alpha_0, \qquad \lambda=0 \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (20)$$

We thus, from this new way of looking at the problem, reach the same conclusion as regards the upper end of the $(\lambda,\ \theta)$ curve as was expressed in equations (11). Regarding the lower end of the curve, we reach the more rational conclusion that the conductivity approaches zero when the water content approaches a certain small finite value.

§18. *Capillary and electrical conductivity of soils.*—Leaving now these rudimentary theoretical considerations concerning the probable manner in which the capillary conductivity depends on the water content, let us see what can be done from the experimental point of view. In considering the capillary potential ψ, we have seen that we may get results from static experiments in which the capillary pull of the soil is balanced against the constant pull of gravity. The corresponding experimental method for finding λ would be to expose a moist soil to

the constant pull of gravity and find how the velocity of the water current depended upon the moisture content. Or we might, knowing already how the potential varied with the water content, keep two ends of a horizontal soil column at such water contents that their difference of potential was constant, and then find how the rate of flow from one end to the other depended upon the mean water content, finally making the gradient infinitesimal. No experiments have been made by either of these methods and the conditions necessary for obtaining trustworthy results appear difficult of realization. We may, however, have recourse to an entirely different and indirect method.

Consider the conductivity of a moist soil for an electric current. If we assume, as is doubtless allowable in most cases at least, that the conductivity of the solid soil grains is negligible in comparison with that of the moisture, which always contains some dissolved electrolytes, we see that the paths by which the electric current passes through the soil are the same as those along which a current of water would flow, the stream lines being identical if, as we do all along, we assume the capillary flow to be so slow that the kinetic energy of the motion is negligible in comparison with the dissipation. Assume, as was justified by the reasoning of section 13, that the films of water over the soil grains remain of sensibly constant thickness so long as the soil is not very dry. Then the conductivity of the films per square centimeter will be constant both for water and for electricity, if the soil has been wet so long that the composition of the soil moisture is not undergoing any further change. A variation of the water content of the soil with its consequent change of the arrangement of the films will now influence the capillary and the electrical conductivities in the same way, and they will be proportional to each other.

Several tacit assumptions are involved in this statement. One of these is that since the films are very thin, the electrical conductance depends sensibly only on the films and not on the capillary water, so long as the capillary water does not form continuous paths through the soil. The analogous assumption regarding the flow of water has already been considered in section 14; it appears legitimate. The second assumption involved is that when more water is added to the soil the composition of the films as regards electrolytes remains constant. This might not be the case with some very insoluble soils, but the difficulty may be obviated by moistening the soil with a salt solution of much greater conductivity than the soil solution but not strong enough to have a very different surface tension. The conductivity of this solution will be sensibly independent of the nature of the soluble substances in the soil. It seems, therefore, as if experiments on the electrical conductivity of moist soils ought to give us valuable information about the capillary conductivity; not, to be sure, about its absolute value, for the ratio of capillary to electrical conductivity might

vary greatly from soil to soil, but regarding the variation with changing water content in any given soil.

We have only very scanty data on this point, as no experiments have been made expressly for this purpose; but we have a few, which may be found on pages 21 and 23 of Bulletin No. 8 of the Bureau of Soils. The experiments were on the electrical resistance of quartz sand wet with solutions of K_2SO_4 and NaCl. In figure 18 the conductivities found are plotted as ordinates against the moisture contents as abscissas. The curves are of an approximately parabolic form which does not contradict our conclusions from theoretical considerations regarding the middle part of the (λ, θ) curves. Another set of results obtained in a somewhat different way, but showing the same thing, is plotted in figure 19, giving a similar curve.

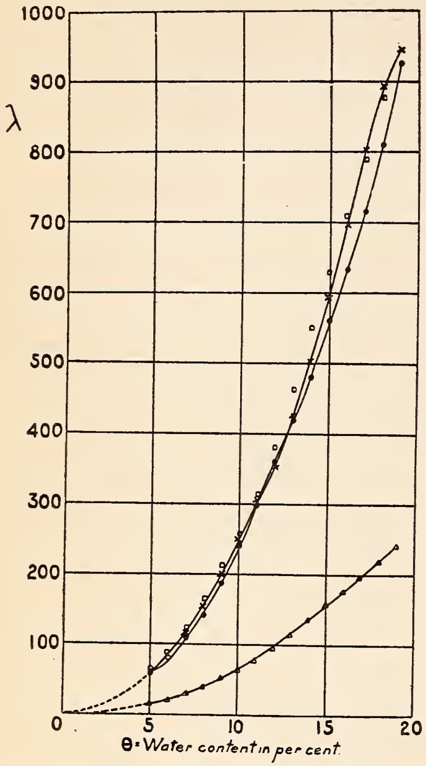

FIG. 18.—Electrical conductivity of moist quartz sand. ⊙ = N/20 K_2SO_4. × = N/10 NaCl. △ = N/40 NaCl. ⊡ = N/40 Na Cl) × 4.

The only other data available are from some experiments on Sassafras sandy loam performed in the Physical Laboratory of the Bureau of Soils by Mr. A. G. McCall. The soil was wet with water without any added electrolyte. The results are shown in figure 20, and while they are somewhat irregular, they indicate the same general type of curve as already found.

IV. THE CAPILLARY FLOW OF WATER.

§ 19. *Nature of the problem.*—As we have shown in section 7, if Q be the capillary current density at a given point in the direction x, if λ be the capillary conductivity of the soil at the given point, and if S be the rate at which the capillary potential increases in the direction x, we may write

$$Q=\lambda S \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (21)$$

or

$$Q=\lambda \frac{\partial\psi}{\partial x} \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (22)$$

or again

$$Q=\lambda \frac{\partial\psi}{\partial\theta} \cdot \frac{\partial\theta}{\partial x} \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (23)$$

If λ and $\frac{\partial \psi}{\partial \theta}$ were constants, this equation (23) would be identical in form with the Fourier-Ohm law, and all the mathematical results obtained from that law would be at once applicable to problems in the capillary flow of water through soils.

Since we have already shown in Section II that $\frac{\partial \psi}{\partial \theta}$ is not constant, and in Section III that λ also is probably not constant, the case is not so simple, and computations have shown that the results of flow experiments can not be represented in this way, in general. If we knew the mathematical forms of ψ and λ as functions of θ, and of θ as a function of x, it is possible, though not probable, that we could give a complete mathematical treatment of the subject. Since we have no such information, we shall turn at once to the consideration of some of the experiments which have been made with a view to finding out how the rate of flow of water from a wet to a dry portion of a given soil depends on the nature of the soil and on the mean water content of the soil.

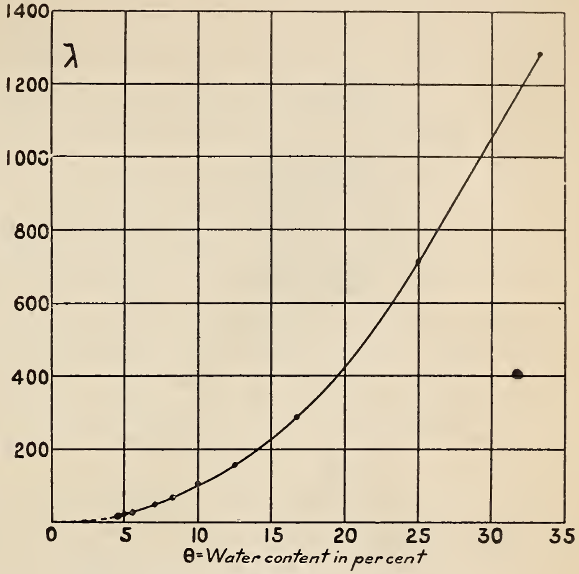

FIG. 19.—Electrical conductivity of moist quartz sand.

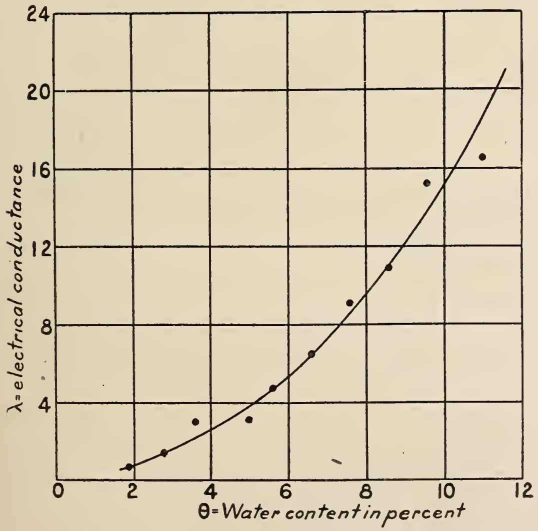

FIG. 20.—Electrical conductivity of moist soil.

§ 20. *Experimental methods.*—Since we are not in possession of the mathematical theory so as to be able to treat states which vary with

the time, the proper method to use seems to be to make experiments on a steady state. If we could arrange to keep one end of a soil column at a given fixed water content and the other at a different fixed water content, and then wait for a steady current to be established, we could, by measuring this rate of flow and determining the final distribution of water in the soil, find how the rate of flow depended on the gradient of water content and on the water content itself. A complete series of experiments of this sort would give us, empirically, complete information about the behavior of the soil in question as a conductor of capillary water. Such attempts as have been made to use this method have hitherto been frustrated by the experimental difficulties, of which the most important are (1) the difficulty of keeping one end of a soil column at a low fixed water content, and (2) the very long time needed for the establishment of the steady state. In addition to these there is the fact, which is always cropping up in physical experiments on soils, that exact duplication of the mechanical condition of granulation and packing is so far from possible as to require averaging data from many experiments before conclusions can safely be drawn.

Recourse has therefore been had to a method much simpler in practice, though more complex in theory. In a Welsbach lamp chimney of about 4.6 cm. internal diameter, and closed at the bottom with a cork or rubber stopper, were placed two layers of soil, each about 3 cm. thick, the upper resting directly on the lower. The two layers had different water contents, and each was uniformly wet at the start. The soil was worked up by hand into a condition of perfect tilth and was lightly tamped when put into the tube. The lamp chimney was closed at the top by another stopper. In the earlier experiments the lower layer was always the wetter, but the later experiments were run in duplicate, each tube with the wet layer below being accompanied by another with the wet layer above. After the two layers of soil had been in contact for a certain time they were separated and the total water content of each determined.

In expressing the results the most natural way would be to measure the movement of water in grams and the water content in either grams per cubic centimeter or per cent of the weight of dry soil. If the mass of dry soil is not the same for each layer, a given motion of water from the wet layer to the dry will not cause equal changes in the water contents measured in per cent. It was found, however, in analyzing the results that such general relations as were to be found at all came out quite as clearly when all quantities of water were expressed in percentage of the dry weight of soil as when they were expressed in more rational units. Hence the results have been expressed in this way because the amount of computation involved is much less.

The initial water contents of the two layers were varied very widely, the wet layer being sometimes too wet to be worked properly and the dry layer being sometimes air dry. Sometimes the initial difference of water content was very large and sometimes only 1 or 2 per cent. In expressing the results the ratio has been taken of M, the mean change in the percentage of moisture in the layers, to $\varDelta$, the initial difference of percentage. Repeated attempts have failed to detect any systematic variation of the value found for this quantity $\frac{M}{\varDelta}$ with the magnitude of $\varDelta$. It does, however, seem probable that $\frac{M}{\varDelta}$ varies systematically with the mean water content of the two layers, which will be denoted by $\bar{\theta}$. In other words, for a given mean water content $\bar{\theta}$, the amount of water M that moves from one layer to the other in a given time is proportional to the initial difference of water content—i. e., to the initial mean gradient of water content—while this motion for a unit initial difference or $\frac{M}{\varDelta}$ is dependent to some degree on the value of $\bar{\theta}$.

The soils used were of various types, as follows:

(1) Garden loam from the grounds of the Department of Agriculture.

(2) Takoma lawn soil—a sandy loam.

(3) Podunk fine sandy loam.

(4) Cecil clay.

(5) Porters black loam—a heavy muck soil.

(6) Leonardtown loam.

(7) Windsor sand subsoil.

(8) New Mexico dunesand.

(9) Sharkey clay.

(10) Susquehanna clay.

(11) Black adobe.

§ 21. *Summary of experimental results.*—The results obtained by the method just described are collected in the following tables. In about half the cases the figures are based on a single experiment, while in the remainder they are based on two.

Tables X and XI contain the figures for garden loam for various periods of contact of the two layers. In these experiments the wet layer was always at the bottom, so that the values of $\frac{M}{\varDelta}$ are somewhat smaller than would have been the case if, as in later experiments, the tubes had been run in pairs, one with the wet layer below and the other with the dry layer below.

TABLE X.—*Flow of water from wet to dry layers of garden loam.*

2 days.		4 days.		9 days.		15 days.	
$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$
3.41	0.250	3.50	0.314	11.05	0.278	11.05	0.287
5.80	0.205	5.78	0.239	12.14	0.282	12.08	0.312
8.31	0.205	8.21	0.237	13.62	0.224	13.38	0.260
10.78	0.200	10.64	0.225	14.62	0.239	14.56	0.248
13.33	0.225	13.58	0.264	15.09	0.308	15.08	0.329
				17.64	0.251	17.42	0.237
Means ..	0.217	....	0.256	....	0.264	....	0.279

TABLE XI.—*Flow of water from wet to dry layers of garden loam.*

2 days.		3 days.		5 days.		7 days.	
$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$
9.68	0.162	9.64	0.197	9.66	0.253	9.67	0.267
10.90	0.179	10.86	0.201	10.86	0.243	10.83	0.273
12.15	0.200	12.17	0.223	12.36	0.256	12.27	0.275
14.11	0.264	14.06	0.272	14.04	0.288	14.03	0.305
Means..	0.201	....	0.223	....	0.260	....	0.280

Tables XII, XIII, and XIV contain figures for Podunk fine sandy loam. In each series of Table XII the initial water content of the dry layer was constant, only the wet layer being varied. Additional data are given in Tables XIII and XIV.

TABLE XII.—*Flow of water from wet to dry layers of Podunk fine sandy loam.*

[Time, 4 days. D_0 = initial water content of dry layer.]

$D_0=0.5$		$D_0=4.8$		$D_0=9.5$		$D_0=13.8$		$D_0=18.1$	
$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$
2.63	0.300	....	....	....	....	....	....	....	....
5.13	0.333	7.21	0.295	....	....	....	....	....	....
8.22	0.380	9.53	0.301	11.63	0.251	....	....	....	....
9.63	0.356	11.80	0.283	13.99	0.236	16.25	0.369	....	....
11.49	0.315	13.41	0.287	15.54	0.237	17.94	0.168	20.10	0.385
14.12	0.289	16.19	0.208	19.00	0.297	20.67	0.360	22.66	0.485
8.54	0.329	11.63	0.275	15.04	0.255	18.29	0.299	21.38	0.435

TABLE XIII.—*Flow of water from wet to dry layers of Podunk fine sandy loam.*

2 days.		4 days.		4 days.		4 days.	
$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$
3.75	0.282	3.76	0.296	0.86	0.388	9.14	0.389
6.07	0 312	6.14	0.337	1.66	0.286	10.00	0.380
8.10	0 321	8.23	0.317	2.47	0 316	10.87	0.383
10.71	0.302	10.58	0.296	3.29	0.331	11.46	0.371
13.54	0.286	13.48	0.285	4.21	0.353	12.49	3.375
				5.62	0.382	13.90	0.373
....	0.301	....	0.306	....	0.343	....	0.378

TABLE XIV.—*Flow of water from wet to dry layers of Podunk fine sandy loam.*

Time in hours.	Observed value of $\frac{M}{\Delta}$	$\bar{\theta}$	Δ_1	Time in hours.	Observed value of $\frac{M}{\Delta}$	$\bar{\theta}$	Δ_1
1	0.20	12.3	13.9	96	0.27	12.9	15.2
2	0.24	12.3	13.9	120	0.27	12.9	15.2
3	0.23	12.8	13.75	168	0.26	12.9	15.2
4	0.24	12.1	13.9	192	0.27	12.9	15.2
24	0.25	13.0	15.5	240	0.28	12.9	15.2
24	0.28	12.9	15.2	288	0.28	12.9	15.2
48	0.27	12.9	15.2	360	0.21	12.9	15.2
72	0.27	12.9	15.2				

In Tables XV and XVI are given data obtained from a less extended series of experiments on eight different soil types, the time being two days for all of them, so that the results are comparable.

TABLE XV.

[Time, 2 days.]

Porters black loam.		Windsor sand subsoil.		Leonardtown loam.		New Mexico dune-sand.	
$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$
26.10	0.131	5.15	0.266	7.90	0.190	2.02	0.274
45.35	0.271	9.42	0.301	15.80	0.270	6.67	0.187
....	0.201	....	0.283	....	0.230	....	0.230

TABLE XVI.

[Time, 2 days.]

Cecil clay.		Sharkey clay.		Susquehanna clay.		Black adobe.	
$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$
8.85	0.217	14.85	0.201	10.72	0.228	11.22	0.214
12.22	0.254	24.12	0.160	18.97	0.249	19.37	0.233
....	0.235	....	0.180	....	0.238	....	0.223

TABLE XVII.

[Time, 2 days; $\frac{N}{2}NH_3$]

Cecil clay.		Sharkey clay.		Susquehanna clay.		Black adobe.	
$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$
8.02	0.213	13.52	0.207	9.95	0.238	9.22	0.144
13.77	0.237	19.70	0.184	15.92	0.201	15.47	0.191
....	0.225	....	0.195	....	0.219	....	0.167

The experiments from which the results recorded in Table XVII were obtained were similar to those of Table XVI, except that the soils were moistened with a half-normal ammonia solution instead of with water.

The results given in Table XVIII were obtained by a slightly different method, the layers of soil being thicker and the analysis for water content having been made not on the whole of each layer but on a 1-inch section adjacent to the bounding surface. The results are comparable with one another, but not with those recorded in the previous tables.

TABLE XVIII.—*Flow of water from wet to dry layers of soil.*

[Time, 13 days; thick layers.]

Garden loam.		Podunk fine sandy loam.		Takoma soil.		Cecil clay.	
$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$	$\bar{\theta}$	$\frac{M}{\Delta}$
....	....	1.29	0.379	....	....	....	....
....	....	2.28	0.351	3.65	0.279	....	....
6.18	0.270	4.91	0.370	5.97	0.252	7.25	0.348
9.87	0.226	8.47	0.323	9.57	0.230	10.03	0.264
14.82	0.281	11.66	0.313	11.50	0.308	14.20	0.324
15.85	0.264	14.10	0.278	13.27	0.305	15.32	0.290
19.05	0.338	18.51	0.303	17.47	0.373	16.22	0.378
24.38	0.305	21.70	0.302	....	....	....	....
Means..	0.281	....	0.327	....	0.291	....	0.321

§ 22. *Comparison of different soils.*—The first thing that strikes one upon looking through the foregoing tables is that the values of $\frac{M}{\Delta}$, or the mean change of water content per unit initial difference, do not vary very much with the conditions of the experiment nor from soil to soil.

In order to compare the results for different soils the arithmetical mean of all the values of $\frac{M}{\Delta}$ obtained for each soil, with a duration of two days, has been taken, and these means are given in Table XIX.

TABLE XIX.—*Mean values of $\frac{M}{\Delta}$ for different soils.*

[Time, 2 days.]

Type of soil.	Mean value of $\frac{M}{\Delta}$	Total number of separate experiments.	Type of soil.	Mean value of $\frac{M}{\Delta}$	Total number of separate experiments.
New Mexico dunesand.......	0.230	4	Sharkey clay..................	0.187	8
Windsor sand subsoil	0.283	4	Susquehanna clay	0.228	8
Podunk fine sandy loam	0.296	12	Black adobe	0.195	8
Garden loam	0.226 (0.209)	9	Porters black loam	0.201	4
Leonardtown loam...........	0.230	4	Mean	0.231	....
Cecil clay	0.230	8			

The values for Garden loam were corrected before averaging to allow for the fact that in this case the wet layer was always below. The uncorrected value is given in parenthesis. The values of $\frac{M}{\Delta}$ vary only from 0.187 to 0.296, one-half the values being very close to the mean, which is 0.231. In the table the light soils are at the top and the heavy soils at the bottom. On the whole, the heavy soils give low values of $\frac{M}{\Delta}$, while the highest values are given by two sandy soils containing particles of a very wide range of sizes. Further experiments with different soil types might, however, fail to confirm this observation.

A second observation is that with four heavy clay soils the flow of water was nearly the same, whether the soil was moistened with tap water or with a half-normal ammonia solution (see Tables XVI and XVIII), the only striking difference being with black adobe, the heaviest of all. This suggests an interesting line of experimentation on the influence of flocculating agents, in the form of commercial fertilizers, on the rate of capillary flow through soils.

§ 23. *Effect of varying duration of the experiment.*—Let us next examine the variations of $\frac{M}{\Delta}$ with the duration of the experiment. In Table XX are collected the mean results of the various, not always strictly comparable, results for garden loam for various durations of the experiment from two to fifteen days. The increase of $\frac{M}{\Delta}$ from the top to the bottom of the table is not large. The change in the corrected values is somewhat more, but the process of correction was a very doubtful one.

TABLE XX.—*Variation of total flow, with the duration of the experiment.*

[Garden loam.]

Duration of experiment in days.	Total number of tubes.	Mean value of $\frac{M}{\Delta}$	$\frac{M}{\Delta}$ corrected for gravity.
2	14	0.212	0.223
3	4	0.223	0.239
4	10	0.256	0.269
5	4	0.260	0.278
7	4	0.280	0.299
9	12	0.264	0.338
15	12	0.279	0.355

Turning back to Table XIV, we have a series of strictly comparable experiments on Podunk fine sandy loam. And here we see that there is no certain increase of $\frac{M}{\Delta}$ from one to fifteen days and that two-thirds

of the change has occurred in the first hour. It will also be seen, in the other cases to be found in the tables, that the variations of $\frac{M}{\Delta}$ with the time are, in general, no greater than the uncertainties of the figures.

It thus appears that the very great initial gradient of water content at the contact of the two layers of soil produces a sudden translocation of water and a readjustment and great diminution of this gradient; and that after a short initial period the further motion is very slow—too slow to be measured satisfactorily by the methods here pursued, unless the duration of the experiments were much increased. It remains a question why this sudden initial change reduces the average difference of water content to about one-half its initial value for all the different soils,[a] as it appears to do.

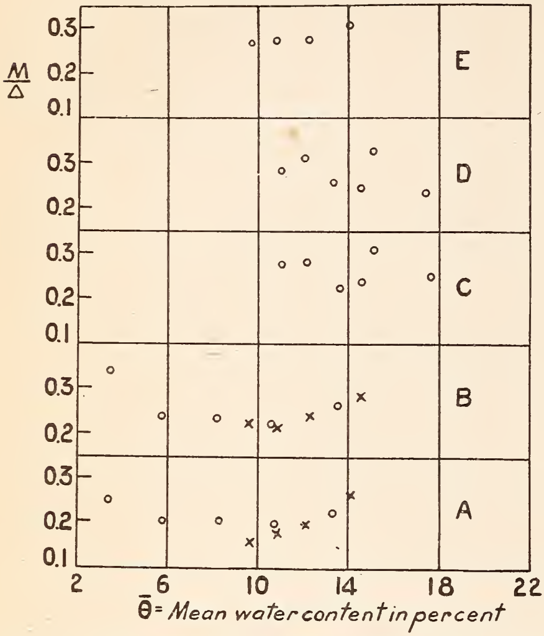

FIG. 21.—Diagram showing changes in mean water content $\left[\frac{M}{\Delta}=f(\bar{\theta})\right]$ in layers of Garden loam. A, 2 days, tables 10 and 11; B, 4 days, tables 10 and 11; C, 9 days, table 10; D, 15 days, table 10, E, 7 days, table 11.

§ 24. *Variation of* $\frac{M}{\Delta}$ *with* $\bar{\theta}$.—The only extended series of measurements are those made upon garden loam and Podunk fine sandy loam. If we examine these, we do not at first see any systematic connection between $\frac{M}{\Delta}$ and the mean water content of the soil, or $\bar{\theta}$. The variations of $\frac{M}{\Delta}$ are, in fact, rather irregular; but if we plot the observations, a certain systematic run suggests itself. Taking the values of $\bar{\theta}$ as abscissas and the values of $\frac{M}{\Delta}$ as ordinates, we get, from the data of Tables X to XIV, the points plotted in figures 21 for garden loam and 22 for Podunk fine sandy loam.

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[a] The final difference of water content is $(\Delta-2M)$; hence, since on the average we have found $\frac{M}{\Delta}=0.23$, we have $\frac{\Delta-2M}{\Delta}=1-0.46$, or about $\frac{1}{2}$.

Looking first at figure 21, we have in the lowest section (A) results obtained from experiments of two days' duration. The points marked ⊙ are from Table XV and those marked × are from Table XI, representing two distinct series of observations. In the next section (B) the points marked ⊙ are from Table X, and refer to experiments of four days' duration. The points marked × are from Table XI, and were obtained by averaging the results for two similar experiments, one of which ran for three days and one for five. In sections (C), (D), and (E) are the results obtained for nine, fifteen, and seven days, respectively. The distribution of the points for the two-day and four-day series, (A) and (B), distinctly indicates a minimum value of $\frac{M}{\Delta}$ for a mean water content $\bar{\theta}$ of about 10 per cent, and the other series, while very irregular, does not contradict this.

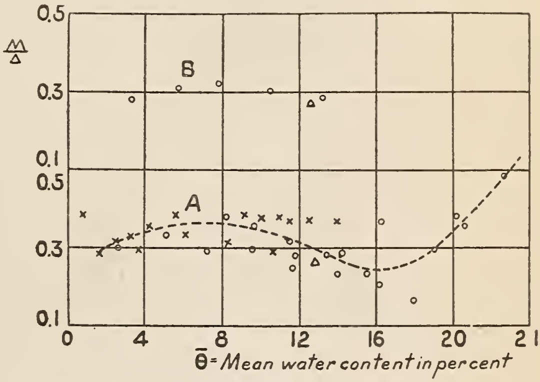

FIG. 22.—Diagram showing changes in mean water content $\left[\frac{M}{\Delta}=f(\theta)\right]$ in layers of Podunk fine sandy loam. A, 4 days, ⊙, table 12; ×, table 13; ⚠, table 14. B, 2 days; ⊙, table 13; ⚠ table 14.

The observations for Podunk fine sandy loam shown in figure 22 are more numerous. In the lower section (A) are plotted all the values obtained from various experiments for a duration of four days, and in the upper section (B) are a few values for two days. The four-day points, while pretty well scattered, seem to indicate that as the mean water content of the soil, $\bar{\theta}$, increases from zero, the rate of flow for unit initial difference of water content, or $\frac{M}{\Delta}$, at first increases; passes through a rather flat maximum at about 6 to 10 per cent mean water content; decreases to a minimum in the vicinity of 16 per cent and thereafter increases rapidly. It would of course be absurd to try to represent all these points by a curve, but the general run of the observations seems to be qualitatively somewhat as shown by the dotted curve of the lower part of the figure. The two-day points, so far as they go, have a somewhat similar distribution.

The existence of a minimum speed of capillary flow at somewhere about one-half to two-thirds of the optimum water content seems to be clearly indicated for these two soils. The observations also indicate that the water content must be reduced to very low values before the speed of flow per unit gradient of water content falls to anywhere near zero if it ever does so.

§ 25. *Possible interpretation of the results of § 24.*—By § 19 we have for the instantaneous value of the current at any point

$$Q=\lambda\frac{\partial\psi}{\partial x} \quad \ldots\ldots\ldots\ldots \quad (22)$$

or

$$Q=\lambda\frac{\partial\psi}{\partial\theta}\cdot\frac{\partial\theta}{\partial x} \quad \ldots\ldots\ldots\ldots \quad (23)$$

Whence

$$\frac{Q}{\frac{\partial\theta}{\partial x}}=\lambda\frac{\partial\psi}{\partial\theta}$$

What we have measured in the experiments just described is the *total* flow M for a given initial *mean* gradient $\varDelta$. It seems probable that the variations of $\frac{M}{\varDelta}$ with θ will be at least parallel to the variations of $\frac{Q}{\frac{\partial\theta}{\partial x}}$. In other words, if we interpret λ as the mean conductivity of the two layers of soil and $\frac{\partial\psi}{\partial\theta}$ as the mean value of $\frac{\partial\psi}{\partial\theta}$ between the limits of water content actually used, it seems that we must have an equation of the general form

$$\frac{M}{\varDelta}=\lambda\frac{\partial\psi}{\partial\theta}\times\text{constant} \quad \ldots\ldots \quad (24)$$

Let us assume that such an equation does hold. Let us also assume that $\lambda=f(\theta)$ has the form already predicted in Section III. Take also the results for Podunk fine sandy loam regarding the relation of $\frac{M}{\varDelta}$ to θ and let us see what we can learn about the probable value of $\frac{\partial\psi}{\partial\theta}$. As we have no adequate numerical data it will suffice to use ideal values.

Suppose that on a diagram with θ as abscissa we plot, first a curve showing $\frac{M}{\varDelta}$ as a function of θ, second, a curve showing λ as a function of θ. Then by equation (24), which may be put in the form

$$\frac{\partial\psi}{\partial\theta}=\frac{\frac{M}{\varDelta}}{\lambda}\times\text{constant} \quad \ldots\ldots\ldots\ldots \quad (25)$$

we may find how $\frac{\partial\psi}{\partial\theta}$ would have to vary with θ in order to satisfy equation (24). The only object in doing this is to see whether the results appear rational or not, when compared with the results obtained in Section II.

In figure 23 are plotted: (A) a curve with the general form suggested by figure 22 and representing, in an ideal case, the relation of $\frac{M}{\Delta}$ to θ; (B) a curve to illustrate a possible form of the relation of λ to θ consistent with the reasoning and the conclusions of Section III; (C) a curve of which the ordinates are ten times the values obtained by dividing the ordinates of A by those of B. This curve (C) therefore shows how $\frac{\partial\psi}{\partial\theta}$ should vary with θ if the forms of $\frac{M}{\Delta}$ and of λ as functions of θ were as shown in curves A and B. These values of $\frac{\partial\psi}{\partial\theta}$ should have their signs reversed as will easily be seen from the fact that what we have called Δ corresponds not to the mean value of $\frac{\partial\theta}{\partial\psi}$ but to the negative of that, Δ being positive when θ *decreases* with increasing x. By examining the form of the curve C for $\frac{\partial\psi}{\partial\theta}$ it will be found that it is not inconsistent with the experimental curves actually found for ψ as a function of θ and given in figures 7 to 13.

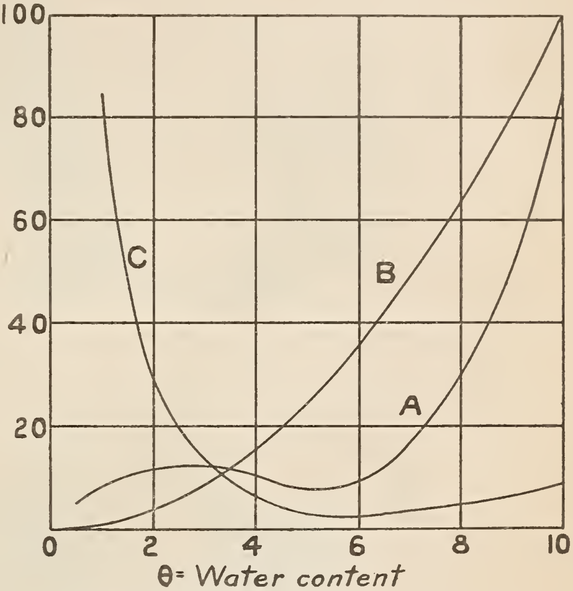

FIG. 23.—Diagram showing theoretical relationship between water content, conductivity for moisture, and the potential function. A, $\frac{\mu}{\Delta}=f\theta$. B, $\lambda=\theta^2$. C, $\frac{\partial\psi}{\partial\theta}=f(\theta)=\frac{10}{\lambda}\cdot\frac{M}{\Delta}$.